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Interaction of the Terrestrial and Atmospheric Hydrological Cycles in the Context of the North American Southwest Summer Monsoon

TRMM



*A Report to the National Aeronautics and
Space Administration
on Project Number NAG 5-1642*

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Final Technical Report
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A. Abstract

Work under this grant has used information on precipitation and water vapor fluxes in the area of the Mexican Monsoon to analyze the regional precipitation climatology, to understand the nature of water vapor transport during the monsoon using model and observational data, and to analyze the ability of the TRMM remote sensing algorithm to characterize precipitation.

An algorithm for estimating daily surface rain volumes from hourly GOES infrared images was developed and compared to radar data. Estimates were usually within a factor of two, but different linear relations between satellite reflectances and rainfall rate were obtained for each day, storm type and storm development stage. This result suggests that using TRMM sensors to calibrate other satellite IR will need to be a complex process taking into account all three of the above factors. Another study, this one of the space-time variability of the Mexican Monsoon, indicate that TRMM will have a difficult time, over the course of its expected three year lifetime, identifying the diurnal cycle of precipitation over monsoon region. Even when considering monthly rainfalls, projected satellite estimates of August rainfall show a root mean square error of 38%. A related examination of spatial variability of mean monthly rainfall using a novel method for removing the effects of elevation from gridded gauge data, show wide variation from a satellite-based rainfall estimates for the same time and space resolution.

One issue addressed by our research, relating to the basic character of the monsoon circulation, is the determination of the source region for moisture. The monthly maps produced from our study of monsoon variability show the presence of two rainfall maxima in the analysis normalized to sea level, one in south-central Arizona associated with the Mexican monsoon maximum and one in southeastern New Mexico associated with the Gulf of Mexico. From the point of view of vertically-integrated fluxes and flux divergence of water vapor from ECMWF data, most moisture at upper levels arrives from the Gulf of Mexico, while low level moisture comes from the northern Gulf of California. Composites of ECMWF analyses for wet and dry periods (classified by rain gauge data) show that both regimes show low level moisture arriving from northern and central Gulf of California. Above 700 MB, moisture comes from both source regions and the Sierra Madre Occidental. During wet periods a longer fetch through the moist air mass above western Mexico results in a greater moisture flux into the Sonoran Desert region, while there is less moisture from the Gulf of Mexico both above and below 700 mb.

Work on the grant subcontract at the University of Colorado concentrated on the development of a technique useful to TRMM combining visible, infrared and passive microwave data for measuring precipitation. Two established techniques using either visible or infrared data applied over the US Southwest correlated with gauges at the 0.58 to 0.70 level. The application of some established passive microwave techniques were less successful for a variety of reason, including problems in both the gauge and satellite data quality, sampling problems and weaknesses inherent in the algorithms themselves. A more promising solution for accurate rainfall estimation was explored using visible and infrared data to perform a cloud classification, which when combined with information about the background (e.g. land/ocean), was used to select the most appropriate microwave algorithm from a suite of possibilities.

B. Work performed

The objectives of our TRMM program have been to:

- analyze the *in-situ* observed precipitation climatology;
- combine atmospheric models with precipitation measurements; and,
- analyze the ability of the TRMM remote sensing algorithm to characterize precipitation.

Research relevant to these objectives is described below. Each description can be pursued in more detail in the appropriate appendix.

1. Estimating Surface Precipitation over Mexico by Calibrating Satellite Infrared Imagery and Airborne Radar (See Appendix I for manuscript)

An algorithm for estimating daily surface rain volumes from hourly GOES infrared images was developed. Using data from the Southwest Area Monsoon Project (SWAMP) undertaken during the summer of 1991, linear relations between digital infrared counts and cloud radar reflectivities were derived with digital IR counts as the independent variable. These relations were applied to hourly GOES IR images to provide grid point reflectivity estimates which were used in conjunction with a regional reflectivity-rainrate (ZR) relation to generate surface rainrate estimates. Daily rain volumes were then calculated using a simple grid point summation, and two area-time integral approaches. Rainfall estimates were generally within a factor of two, while comparison of the relative performance of the three estimation techniques was inconclusive.

The linear relations among the cases employed in the study were determined through a regression analysis between digital IR counts and airborne internal radar reflectivity samples. This revealed a positive relation between the two in all samples analyzed, where stratiform samples displayed a higher degree of correlation than did convective types. It was also found that each linear relation was unique to the day, type of storm, and stage of storm development. This result suggests that any calibration of satellite IR rainrates (e.g. GOES GPI) done via TRMM sensors will need to be a complex process, taking into account these three variables.

2. Space-Time Variability of the Mexican Monsoon (See Appendix II for draft manuscript)

Rainfall from the Mexican monsoon is difficult to measure or interpret because it is particularly variable in time and space. Monthly and hourly long-term rain gage records were used to quantify rainfall variability in terms of the diurnal cycle, frequency distribution of monthly rainfall, and the spatial persistence of monthly rainfall anomalies. The diurnal cycle and form of the monthly frequency distribution were found to vary spatially within the area affected by the Mexican monsoon. Gage data were also used to construct a 30-year sequence of hourly precipitation averaged over a 4 degree by 5 degree grid (and percent area receiving rain).

These areal data were used to compare point versus areal characteristics of rainfall and to provide the basis for satellite sampling experiments. The areal rainfall was sampled once every 13 hours in order to estimate temporal sampling errors of rainfall measured with the TRMM satellite. Three years of "satellite" sampling were found to be insufficient for identifying the diurnal cycle. The root mean square error of "satellite" estimates of August rainfall was 38%. This variability in monsoon precipitation indicates that, at least for this region and likely for other arid lands, TRMM over the short term will be insufficient to provide a complete precipitation climatology. This shortcoming may be mitigated if further work is done to find ways of combining optimally TRMM rainfall measurements with ground-based radar and gauge measurements.

3. Spatial and Elevational Variations of Summer Rainfall in the Southwestern United States (see Appendix III for manuscript)

This study examines the spatial variability of mean monthly summer rainfall in the southwestern United States with special attention given to the effect of elevation. Rain gage data from a consistent 60-year period show that mean rainfall increases linearly with elevation within a local area. A simple model (rain = normalized rainfall as a function of latitude and longitude + elevation coefficient * elevation) explains a large part of the spatial variability of mean rainfall. The rainfall model (the MSWR model) and digital elevation data were used to produce a 1° x 1° gridded rainfall climatology for July, August, and September. Regional rainfall estimated with this model is 9.3% higher than an estimate based on arithmetic averaging of gage data over 2° x 2° areas. For individual 2° x 2° cells, the difference between model rainfall and the arithmetic mean of gage rainfall ranged from -250% to +41%.

The MSWR model was used to remove orographic effects from regional rainfall fields. When rainfall is normalized to sea level, two rainfall maximums emerge: one in south-central Arizona associated with the Mexican monsoon maximum and one in southeastern New Mexico associated with the Gulf of Mexico. Detrended block kriging (using the MSWR model as an estimate of the long-term trend) and monthly rain gage data were used to produce unbiased areal rainfall estimates which were compared to 1° x 1° satellite-based rainfall estimates. On a month-by-month basis, there were large differences between the two estimates, although the comparison improved after temporal averaging.

4. Water Vapor Transport Associated with the Summertime North American Monsoon as Depicted by ECMWF Analyses (see Appendix IV for manuscript)

The origins and transport of water vapor into the semi arid Sonoran Desert region of southwestern North America are examined for the July-August wet season. Vertically-integrated

fluxes and flux divergences of water vapor are computed for the 8 summers 1985-1992 from ECMWF mandatory-level analyses possessing a spectral resolution of triangular 106 (T106).

The ECMWF analyses indicate that transports of water vapor by the time-mean flow dominate the transports by the transient eddies. Most of the moisture at upper-levels (above 700 mb) over the Sonoran Desert arrives from over the Gulf of Mexico, while most moisture at low-levels (below 700 mb) comes from the northern Gulf of California. There is no indication of moisture entering the Sonoran Desert at low-levels directly from the southern Gulf of California or the tropical East Pacific. Water vapor from the tropical East Pacific can enter the region at upper-levels after upward transport from low-levels along the western slopes of the Sierra Madre Occidental mountains of Mexico and subsequent horizontal transport aloft.

The T106 ECMWF analyses, when only the mandatory-level analyses are used, do not possess sufficient precision to yield accurate estimates of highly differentiated quantities such the divergence of the vertically-integrated flux of water vapor. Even at a T106 resolution, the northern Gulf of California and the terrain of the Baja California peninsula are not adequately resolved.

5. Intraseasonal Variability of the Summertime North American Monsoon (See Appendix V for manuscript)

Intraseasonal variations associated with the North American Summer Monsoon are investigated. Composite wet and dry periods during July and August of 1985-1992, defined from rain gauge data for southeast Arizona, are compared. Cloud top temperature (CCT), horizontal and vertical velocities, specific humidity, precipitable water (PW), convective indices, moisture flux, and parcel trajectories are all examined. ECMWF mandatory-level analyses possessing a spectral resolution of triangular 106 are employed.

Significant differences exist between wet and dry conditions over the Sonoran Desert for all fields considered. As the monsoon shifts from dry to wet conditions, the subtropical ridge shifts $\sim 5^\circ$ latitude toward the north, and PW increases by as much as ~ 1.2 cm (~ 0.5 inches). Parcels in the middle troposphere ascend into the region from the southeast, and the atmosphere becomes more unstable. The result is a significant increase in the frequency of deep convection, as determined from CTT $< -38^\circ\text{C}$.

During both monsoon regimes, most of the water vapor entering the Sonoran Desert at low-levels (below 700 mb) arrives from over the northern and central Gulf of California, with a slightly greater flux into the region occurring during the dry phase. Above 700 mb, moisture transported into the Sonoran Desert during both regimes is a mixture of water vapor from over the Gulf of Mexico and Gulf of California, and from residual convective inputs over the Sierra

Madre Occidental mountains of Mexico. During wet periods, however, a longer fetch through the moist air mass above western Mexico results in a greater moisture flux into the Sonoran Desert aloft. Less water vapor from over the Gulf of Mexico flows into western Mexico and the Sonoran Desert under wet conditions than during dry phases, both above and below 700 mb.

6. University of Colorado Subcontract — Passive microwave and visible/infrared satellite estimation of rainfall: Analysis of the North American monsoon and algorithms for TRMM

Local AVHRR data covering the area of the North American monsoon were used to generate monthly rainfall for the months of June through September for 1991 and 1992, using a visible-only technique (the Highly Reflective Cloud method [HRC] developed by Kilonsky and Ramage [1976]) and an infrared-only technique (the Convective Stratiform Technique [CST] developed by Adler and Negri [1987]), both modified for application to AVHRR data. SSM/I data from Marshall Space Flight Center (MSFC) was used to compute monthly rainfall from passive microwave measurements for the same time period, using two different algorithms, one detailed in Adler et al (1991) and the other described in Ferriday and Avery (1994). The monthly totals were all averaged spatially into two-degree bins for the Southwestern United States and compared to similarly averaged rain gauge amounts reported by the National Climatic Data Center (NCDC). The HRC and CST techniques were found to be highly correlated with one another, with a linear correlation coefficient of approximately 0.91, as both methods focus almost exclusively upon deep convective rainfall. The correlations between these methods and the gauges, however, were much lower, 0.58 for the CST method and 0.70 for the HRC technique. The SSM/I methods were even less correlated, and were not well correlated with each other. It was noted that there are a great many factors which could have caused the problems observed. Passive microwave techniques over land have seen less development and usage in the scientific community than ocean-only algorithms. Additionally, the utility of rain gauges as reliable validation data has yet to be firmly established by consensus. Also, at the time these comparisons were carried out, there was insufficient time to perform rigorous quality control on the SSM/I data. Perhaps most importantly, however, is the temporal sampling issue. Both the NOAA satellites which carry the AVHRR and the DMSP satellites which carry the SSM/I are polar orbiters in sun-synchronous orbits, meaning each one samples a given location on the Earth twice a day at best. If the time of this overpass is not synchronous with the diurnal precipitation cycle, it will be virtually impossible to arrive at a sensible monthly estimate over a short period of time. The HRC and CST methods, while physically indirect and biased toward deep convection, used data from the late afternoon, when there is often considerable rainfall activity occurring in the Southwestern United States, while the SSM/I data was gathered at times later in the evening and early in the morning, when precipitation will be fairly sparse and uncorrelated with its daily maximum.

TRMM will attempt to address this issue by utilizing an orbit which is slightly off of sun-synchronous, thus sampling an entire day over a period of several days for any given location in the tropics. Temporal sampling will still continue to be a serious challenge for any rainfall retrieval methods which use low-orbiting satellite data. We feel that the current focus on rainrate retrieval algorithms should be placed upon what happens on the instantaneous level, and that this is where the strengths of visible/ir and passive microwave data should be combined. This has led to the concept of a combined algorithm for TRMM which would use the visible and infrared data to perform a cloud classification field; the cloud type at a given location corresponding to a passive microwave scan point would then be used to draw a conclusion about the type of precipitation occurring, and then this information would be combined with a statement about the background (i.e. land, water, or coastline) to decide on which microwave algorithm should be used for the computation of the rainrate. Thus the combined approach would include an automated cloud classifier and a suite of pre-existing microwave algorithms, each one tuned for a specific type of situation, such as warm rain over ocean or deep convection over land. In theory, this should enhance accuracy and a sense of physical reliability at the instantaneous level, which would open the door for more accurate monthly estimates.

During the final year of funding from the University of Arizona, we began development of such an algorithm. An automated neural network cloud classifier was built for use with AVHRR data over both land and ocean, and this classifier is now in the final stages of testing before it is applied to a study over the Western United States. This study will compare cloud classifications at 32x32km resolution with radar rainrates at 8x8km resolution supplied by MSFC; these radar rainrates are produced as 15-minute averages by the National Weather Service. A study of the relationship between cloud type and rainrate will be conducted with these data, possibly even leading to a new algorithm which uses cloud type frequency to infer monthly rainfall. Additionally, manually classified GMS data and SSM/I data from the TOGA COARE domain are being used to investigate the performance of a sample combined algorithm relative to several individual algorithms, using the same five case studies which were used for the Global Precipitation Climatology Project (GPCP) 3rd Algorithm Intercomparison Project (AIP-3). Radar rainfall data and processing software were provided for this study by Beth Ebert at the Bureau of Meteorology Research Centre in Melbourne, Australia.

It is hoped that the cloud classifier will form the basis for future work in cloud classification, as an automated classifier which works over all temperate latitudes and either a land, snow, or ocean background would be extremely useful for many different purposes. The full combined algorithm concept cannot be tested with real data until TRMM is in orbit, as the current NOAA and DMSP satellites do not offer many instances of coincident coverage, however the preliminary study done here should give an idea of the utility of such an algorithm.

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Miller, S.W. and W.J. Emery, 1995: The relationship between cloud type and rainrate for a portion of the western United States. In preparation.

Miller, S.W. and W.J. Emery, 1995: A combined cloud classifier and passive microwave rainrate algorithm for use with TRMM data: a preliminary study. In preparation.

Schmitz, J. and S. Mullen, 1995: Intraseasonal variability of the summertime North American monsoon. Submitted to *J. Clim.*

Schmitz, J. and S. Mullen, 1995: Water vapor transport associated with the summertime North American monsoon as depicted by ECMWF analyses. Submitted to *J. Clim.*

APPENDIX I — *Estimating Surface Precipitation over Mexico by Calibrating Satellite Infrared Imagery and Airborne Radar*

ESTIMATING SURFACE PRECIPITATION OVER MEXICO BY CALIBRATING
SATELLITE INFRARED IMAGERY AND AIRBORNE RADAR

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ABSTRACT

An algorithm for estimating daily surface rain volumes from hourly GOES infrared images was developed. Using data from the Southwest Area Monsoon Project (SWAMP) undertaken during the summer of 1991, linear relations between digital infrared counts and cloud radar reflectivities were derived with digital IR counts as the independent variable. These relations were applied to hourly GOES IR images to provide grid point reflectivity estimates which were used in conjunction with a regional reflectivity-rainrate (ZR) relation to generate surface rainrate estimates. Daily rain volumes were then calculated using a simple grid point summation, and two area-time integral approaches. Rainfall estimates were generally within a factor of two, while comparison of the relative performance of the three estimation techniques was inconclusive.

The linear relations among the cases employed in the study were determined through a regression analysis between digital IR counts and airborne internal radar reflectivity samples. This revealed a positive relation between the two in all samples analyzed, where stratiform samples displayed a higher degree of correlation than did convective types. It was also found that each linear relation was unique to the day, type of storm, and stage of storm development.

1. Introduction

Recent studies of the southwestern United States and Mexico have revealed that weather in this region is characterized by two distinct climatological regimes. From October through May the mean middle and upper level tropospheric flow is typically from the west, with precipitation resulting primarily from cyclonic storm systems originating over the Pacific ocean and migrating eastward. During the summer season, June through September, a pronounced 'monsoonal' reversal of the mean winds occurs such that easterly or southeasterly winds now describe the 700 mb to 500 mb streamlines. Concurrently, the dominant precipitation mechanism becomes more convective in nature (Smith and Gall 1989).

Ranging in size and organization from individual and short lived multicell thunderstorms to meso-alpha scale convective complexes, these convective systems produce an annual summertime rainfall maximum. As key components of monsoon rainfall in the southwestern United States and Mexico, these storms play a vital role in the local hydrological cycle. Unfortunately, due to the complex orography of this region and inadequate meteorological observations, quantitative and temporal measurements of surface precipitation are difficult. As a result, other means for determining surface rainfall are desirable.

This paper will present a technique for estimating daily

surface rainfall from hourly infrared satellite images. This will be accomplished by employing separate convective and stratiform linear relations to estimate cloud radar reflectivities from digital infrared counts of GOES images. The resulting reflectivities are then used in conjunction with a local reflectivity-rainrate(ZR) relation to estimate surface rainfall. Daily volumetric rainfall totals are then determined using a simple grid point summation and two area-time integral approaches.

An additional topic of this paper focuses on the uniqueness of the linear relationship derived for the internal reflectivity estimation. Are the relationships independent of the convective or stratiform nature of the cloud observed? Or do they depend on factors such as cloud type, type of storm, and time of day?

2. Data

During the Southwest Area Monsoon Project(SWAMP), internal reflectivity data for several mesoscale convective systems(MCS's) were obtained using the National Oceanic and Atmospheric Administrations P-3 Orion aircraft. Reflectivity measurements were taken by the P-3's vertically scanning, 3.2 cm tail radar using a Forward-Aft Scanning Technique(FAST). Reflectivity data for three MCS's observed during SWAMP are used for this study. Since the principal mesoscale systems of the summer monsoon have been

identified as individual and short lived multicell thunderstorms and meso-beta and meso-alpha scale convective complexes, a MCS of each of these types was chosen for this study.

Each case contains three P-3 airborne radar reflectivity samples. Due to the X-band operating wavelength, the maximum unambiguous radar range was limited to 76 Km, and therefore, complete instantaneous observation of storm wide internal radar reflectivities was impossible. As a consequence, each reflectivity sample corresponds to a given region of the cloud where sample size depends on the duration of the observation and the flight track of the aircraft. In this analysis, all radar samples consist of sixteen horizontal 225×225 km² grids spanning the vertical depth of the scanned region. Horizontal resolution in each grid is 3 Km. Figure 1 depicts vertically averaged horizontal reflectivities for one convective and one stratiform sample. Table 1 lists each sample, type of storm, time of observation, and region of storm observed.

The convective or stratiform nature of each cloud region was determined by the SWAMP in-flight observers using a technique presented in Watson et al.(1988). This method distinguishes between convective and stratiform cloud types through consideration of the areal extent, DBZ thresholds, and overall longevity of the echo patterns displayed on a radar screen.

Along with internal reflectivity data, this analysis employed hourly infrared GOES images with approximately 8 km horizontal resolution. Here, 480x480 km² subsectors of full sector GOES images were extracted for the purpose of isolating the individual mesoscale systems described earlier. Since the reflectivity samples described above corresponded to time increments lying between the hourly GOES intervals, these images were linearly interpolated to correspond to the median time of the radar observation. This was felt to provide a more representative infrared sample with which to perform the correlation and regression analysis. Figure 2 presents the interpolated infrared images to be used for each of these systems.

In addition to the reflectivity and satellite data described above, SWAMP employed an extended surface rain gauge network over Sonora and along the west coast of the Sierra Madre Occidental. Daily rainfall totals are available for two hundred sites in Mexico and are used in the estimate of daily rain volumes.

3. Method of Analysis

In order to estimate rainfall over Mexico from satellite infrared imagery, this analysis consisted of three components. The first aspect focused on establishing positive linear relationships between the digital IR counts and internal reflectivities of the grids listed in Table 1. The

second aspect compared the nine regression equations resulting from the first phase of the analysis. Were these relationships unique for each grid pair, or could some of them be considered the same? Finally, the third component of the analysis concentrated on estimating daily rain volumes for selected regions in Mexico using the regression equations derived previously and a local ZR relationship.

a. Correlation and Regression Analysis

Utilizing satellite infrared and airborne reflectivity data for the three MCS's described earlier, correlations between digital IR counts and vertically averaged horizontal radar reflectivity values were analyzed. This was accomplished after first mapping the 8 km resolution infrared images into the 3 km resolution reflectivity space by way of a Cressman weighting scheme.

Next a computer maximization of the correlation coefficient was performed. This involved positioning the DBZ image at several different places within the 3 km resolution satellite infrared image, calculating the correlation coefficients, and selecting the optimal position as the one with the highest correlation coefficient. The necessity of the grid realignment stems from combined P-3 and satellite navigational errors which make collocation by center point latitude and longitude insufficient for correct alignment of the DBZ and IR reference frames (Schlatter 1986).

Lastly, a linear regression between internal radar reflectivities and digital IR counts was performed. This provided a linear IR-DBZ relationship for each radar sample listed in Table 1. The resulting regression equations are of the form illustrated by (1), where mean values are indicated by over-bars.

$$DBZ = Slope \times (IR - \overline{IR}) + \overline{DBZ} \quad (1)$$

With digital IR counts as the independent variable, these equations make it possible to estimate internal radar reflectivities from infrared pixel values of corresponding satellite images.

b. Regression Line Combination

Assuming a positive linear relation between digital infrared counts and internal reflectivity values was to be found, the regression lines for each sample were compared by case and type(convective vs. stratiform) using a method described by Brownlee(1965). This analysis attempted to determine: 1)whether the regression lines of each case may be combined into one regression line; 2) if regression lines of a common type within a case may be combined; 3) if the regression lines representing stratiform or convective types may be combined across cases; and 4) if all regres-

sion lines may be combined into one regression line.

c. Daily Rainfall Estimation

The final phase of this analysis involved estimating daily surface precipitation over a 720×720 km² area centered over each storm used in the analysis. Before generation of rain fields could begin, it had to be determined which grid points were actively precipitating and whether this rainfall was convective or stratiform in nature. This was accomplished using the method of Adler and Negri as employed in their satellite infrared technique for estimating tropical convective and stratiform rainfall(1987).

After determining precipitating convective and stratiform grid points, surface rainfall was estimated using the derived regression equations. These were applied to the hourly GOES infrared images falling within the 24 hour period of daily surface rainfall observations. This produced hourly horizontal reflectivity fields representing the vertically averaged horizontal reflectivities of the clouds in question. In order to produce the corresponding surface rainfall distributions, the estimated convective reflectivities served as input to $Z = 55R^{1.6}$ and the estimated stratiform reflectivities served as input to $Z = 200R^{1.6}$. These are the local ZR relations used by the National Weather Service in Tucson(National Weather Service, 1982). Daily rainfall totals for the selected areas were

computed from the hourly rainrate fields using three different rainfall estimation techniques.

Since the hourly rain volume assigned to any grid point was simply the product of the rainrate(mm hr^{-1}), the area of the grid square(mm^2), and the length of the precipitation interval(1 hr), the first estimation technique will produce a daily volumetric rainfall amount by simply integrating the hourly rain volumes over all grid points and hourly time intervals. No consideration is given to the overall echo area. This method will henceforth be referred to as the grid point approach.

The second rainfall estimation technique, better known as the Area-Time Integral(ATI) approach, is based upon the theory that total volume rainfall can be estimated from the lifetime measurements of areal rainrate distributions from an individual convective storm or from the instantaneous areawide rainfall distribution from a multiplicity of convective storms. This implies similar rainrate distributions for storms of a particular climatic regime, such that the probability density function of the rainrates, $P(R)$, or the percentage of the area where the rainrate is between R and $R+dR$, is approximately constant.

Using these ideas, Atlas et al.(1990) showed there was an inherent climatological relationship, $S(\tau)$, between volume rainfall, V , and the ATI, where the ATI is the product of the average area with precipitation greater than or equal

to an empirically determined threshold rainrate, τ , and the length of the averaging interval, T . This is depicted by (2). Additionally, it was shown that both sides of (2) can be divided by the total precipitation area giving (3), which relates the instantaneous areawide rainrate, $\langle R \rangle$, and the fraction, $F(\tau)$, of the total precipitation area with rainrates greater than or equal to τ . In both expressions, $S(\tau)$ is the climatological rainrate for the regime divided by the relative frequency with which $\langle R \rangle \geq \tau$. Note, $\langle R \rangle$ will be constant for a particular radar sample such that $S(\tau)$ and $F(\tau)$ will be different for each τ , hence the subscripts in (2) and (3).

$$V = S(\tau_i) \times [A(\tau_i) \times T] = S(\tau_i) \times AT_i \quad (2)$$

$$\langle R \rangle = S(\tau_i) \times F(\tau_i), i = 1, 2, \dots, n \quad (3)$$

$S(\tau)$ and τ can be determined through a linear regression of $\langle R \rangle$ and $F(\tau)$ pairs from several radar scans, where $S(\tau)$ is chosen as the slope of the regression line with the highest correlation and τ is the corresponding rainrate threshold. Or, if the observed area is large enough to contain a sample of rainrates representative of the true popu-

lation distribution, one can determine the climatological rainrate using the pdf as illustrated by (4).

$$S(\tau) = \frac{\int_0^{\infty} rP(r) dr}{\int_0^{\infty} P(r) dr} \quad (4)$$

In this analysis, however, neither lifetime observation nor wide area coverage were available so $S(\tau)$, and τ were determined from rainfall distributions estimated from three random radar snapshots taken within each of three distinct convective systems. It was felt that these random high resolution images, when considered together, contained enough cells in varying stages of development to provide a representative pdf for the rainrates. As a means for determining the validity of this application of the ATI, a comparison of the slope parameters obtained through linear regression of $\langle R \rangle$ and $F(\tau)$ pairs from each DBZ image, and that determined from (4) will be made. If the two slope parameters are approximately equal then it will be assumed that these radar samples provide an accurate representation of the climatological rainrate distribution. Otherwise, they are inadequate.

The final method of estimating volumetric rainfall is

better known as the Height-Area Rainfall Threshold(HART) method. This technique is similar to the area-time integral approach, but in this scheme, the efficiency of precipitation mechanisms within a particular climatic regime is taken into account. Computation of the slope parameter no longer results from the regression of $\langle R \rangle$ against $F(\tau_i)$, but instead is derived through a regression of $[E_e(\tau_i) * \langle R \rangle]$ on $F(\tau_i)$, where $E_e(\tau_i)$ is the effective precipitation efficiency for each rainfall threshold. This parameter tends to reduce $S(\tau)$, and thus, volumetric rainfall in accord with decreased surface precipitation resulting from evaporation, entrainment, and mixing(Rosenfeld et al. 1990). Due to the unavailability of information needed to compute the efficiencies on a grid point basis, the precipitation efficiencies used in this analysis were those derived for summer in Big Springs Texas and presented by Rosenfeld et al.(1990). The choice of the Texas precipitation efficiencies over the GATE efficiencies was based on the proximity of Texas to Mexico, and the fact that both of these climatic regimes are thought to be influenced by moisture fluxes from the Gulf of Mexico during the summertime monsoon period.

4. Results of Correlation and Regression Analysis.

As part of the correlation and regression analysis, the IR and DBZ grids were moved relative to each other until the correlation coefficient was maximized. This assumed a

positive relationship between the two and was felt to correct the inherent positioning errors resulting from incorrect satellite and P-3 navigational information. Optimal IR and corresponding DBZ grids for one convective and one stratiform sample are illustrated in Figure 1. Statistical results of the correlation and regression analysis are presented in Table 2. The corresponding regression equations are found in Table 3 and illustrated in Figure 3.

Before formal discussion of the statistical results, two additional aspects of the analysis must be mentioned. Since the internal reflectivity fields were spatially limited by the wavelength of the P-3 tail radar and the period of observation, some points in the reflectivity grid had DBZ values equal to zero. These points were not used in the analysis as it was uncertain whether the zero value resulted from attenuation, lack of echo, or lack of radar observation. Consequently, since the IR and DBZ images were on identical grids once correlation began, the statistics listed in Table 2 are those computed from grid points with non-zero reflectivities.

Another question requiring consideration was that of degrees of freedom. The assignment of degrees of freedom depends on the number of independent pieces of information available for the decision making process. For a typical correlation or regression analysis it is assumed that values entering into the computations are independent, and

therefore, the number of degrees of freedom are equal to the number of points utilized minus one or two, respectively. In this analysis the determination of degrees of freedom was not so trivial, as spatial correlations existed within both the IR and DBZ grids and each had different horizontal resolutions. Computation of the degrees of freedom utilized in this analysis are discussed in Appendix. 1. The number of degrees of freedom for each sample are presented in Table 2, those numbers on the left of the column are the estimates for the IR grid, while those on the right are the DBZ grid estimates.

Recalling that the convective or stratiform nature of each sample was determined by in-flight observers using on-board radar, it appears that stratiform samples, in general, have lower IR and DBZ variances and covariances than convective samples. Accordingly, stratiform samples displayed larger correlation coefficients and regression line slopes than the convective types. This was expected due to the greater horizontal homogeneity displayed by both infrared images and radar reflectivity fields of stratiform clouds. Convective regions, on the other hand, display considerable infrared topography while their internal reflectivity fields are generally a composition of radar echoes of varying heights, vertical extents, and magnitudes (Smull and Houze 1985, 1987a, 1987b; Gamache and Houze 1982, 1985).

Although digital IR counts and DBZ's of stratiform clouds showed higher correlations than those of convective clouds, the strength of these strong linear relationships varied widely. Squaring the correlation coefficients showed that in the best stratiform case 66% of the variation of IR values was accounted for by the linear relationship with DBZ values, while only 34% was explained in the best convective case. In the worst cases only 10% and 12% of the variance was explained for stratiform and convective samples, respectively.

Additional support for these conclusions comes from the scatter diagrams of Figure 3. These diagrams illustrate the strength of the linear relation through the compactness and linearity of the points. The stronger the linear relationship, the greater the tendency of the points to lie along a straight line, and the smaller the spread of the points about this line. The obvious feature of Figure 3 is the distinct difference in the nature of the convective and stratiform distributions. In the stratiform clouds, the linear relationship was most evident, with the spread of the points increasing as the correlation coefficient decreases. This was also apparent for the convective samples which displayed a much wider scatter and accordingly, a weaker linear relationship. In fact, the distribution of points pertaining to July20A and July20B display an almost random character. The wider scatter evident in the

convective samples was felt to result from the higher spatial variability found within both the IR and DBZ fields of these types.

Along with the general trends one must also consider the anomalies, as found in the samples of July12A and July17C. Although the in-flight observation technique mentioned earlier labeled July12A as convective, the statistics of this sample are more characteristic of the stratiform variety. Similarly, July17C was labeled as stratiform, but statistically it looks to be convective. To account for these anomalies it is speculated that perhaps these samples are neither convective nor stratiform in nature and may represent regions of the MCS's where convective clouds were undergoing a transition to a more stratiform nature. The premise for this assertion arises from the observation by Smull and Houze(1987a) of decreased low level reflectivity in a transition zone coupling the convective and stratiform regions of organized MCS's. Decreasing low level reflectivity would reduce the vertically averaged reflectivity and change the nature of the corresponding linear relationship. Although, the average DBZ for July17C was consistent with other samples labeled as stratiform, this phenomena may account for the reduced average internal reflectivity of the July12A case.

Since positive correlation coefficients and slopes by themselves imply that high digital IR counts are analogous

to high radar reflectivities, it remained to be determined whether or not these positive linear relationships have any statistical significance. Even though all correlation coefficients and regression line slopes were positive, statistical significance at the 95% level was found in 3 out of 4 stratiform, and 3 out of 5 convective samples. Those samples showing nonsignificant linear relations included July17B, July17C, and July20B. As a result, it appears one can reasonably assume that high digital IR counts correspond to high internal reflectivities in stratiform regions of tropical MCS's. For convective regions this assumption seems less valid. In fact, if we relabel July17C as convective, and July12A as stratiform, then 4 out of 4 stratiform samples exhibit significant positive relationships, while only 2 out of 5 convective relations are significant.

Furthermore, the two significant convective relationships, July17A and July20A, represented multicell and meso-beta MCS's, respectively, in the early stages of their lifecycles when individual convective elements are relatively distinct. The nonsignificant convective samples, on the other hand, represented regions where anvils had begun merging so that individual convective elements were harder to discern. As a result, shallower, low level convective elements were obscured by high level cirrus and thus, were undetectable by infrared sensors. The P-3 Radar, however,

could penetrate the cloud and see these hidden cells. In light of this, it seems reasonable that radar echoes and infrared images display significantly different spatial patterns in convective regions of storms approaching the mature stage of development.

5. Results of Regression Line Combination

Of particular interest in this analysis was the possibility of combining the regression line of one stratiform (convective) sample with other stratiform (convective) samples both within and across cases. The ability to combine regression lines from different sample environments would imply similar IR and DBZ relationships and thus, a climatological relationship between these observations. In the same manner, one could infer a daily climatological relationship existed if regression lines of both convective and stratiform samples taken from the same storm can be combined. Synthesis of all samples of all storms would imply the existence of one regional infrared and internal reflectivity relationship. To evaluate each of these possibilities the regression lines being considered were put through a series of hypothesis tests presented in Brownlee(1965) and outlined earlier. Failure of any of these tests would lead to immediate conclusion that these lines can not be combined in a statistical sense.

Overall, all attempts to combine regression lines met

with failure and as a result, it appears that any relationship found between digital IR counts and vertically averaged horizontal reflectivities is tenuous at best. Although this might be anticipated for the highly dynamic convective regions, the lack of any consistent relation among the more benign stratiform types was somewhat surprising. Clearly, the different morphology within each system, ranging from an unorganized multicellular cluster to a highly organized meso-alpha scale convective complex, was sufficient to produce considerable variations among the internal structure of each storm. This ultimately affected radar reflectivity and any relation it may have had with digital IR counts. The lack of any consistent relation between the two has also been noted by Reynolds and Smith(1979).

6. Results of Daily Rainfall Estimation

The final aspect of this analysis involved the estimation of daily precipitation volumes. To estimate precipitation using the methods described earlier, surface rainfall distributions for hourly GOES images had to be created. This was accomplished using three different pairings of convective and stratiform regression equations to generate internal reflectivity fields from which hourly rainfall distributions were produced.

First of all, rainfall distributions were estimated from hourly GOES infrared images using the convective and

stratiform regression equations on a case by case basis. This meant that only regression relations for 12 July 1990 were used to estimate surface rainfall on that day, where a convective(stratiform) relation was applied at grid points labeled convective(stratiform) using the technique of Adler and Negri(1987). In the case of duplicate regression equations(i.e. two representing the same cloud type), the one providing the highest correlation coefficient was used. As a result, the case specific equations used in the rainfall estimation on 12 July 1990 were (12a) and (12b), those for 17 July 1990 were (17a) and (17b), while those for 20 July 1990 were (20a) and (20c).

The second volumetric rainfall estimation was performed using the stratiform and convective regression equations which had the highest correlation coefficients. In this case, (17a) was used to estimate radar reflectivity at all convective grid points and (12b) was used for all stratiform pixels.

Although earlier statistical analysis suggested otherwise, the final pairing involved two regression equations resulting from combination of the two best convective and two best stratiform relations. Accordingly, (17a) and (17b) were combined for use at convective grid points, while equations (12b) and (12c) were combined for use at stratiform grid points.

Before rainfall estimation could begin, rainrate thresholds and slope parameters had to be determined for both the ATI and HART approaches. The regression of $\langle R \rangle$ against $F(\tau_i)$ across all reflectivity images yielded an effective precipitation threshold, τ , of 5 mm hr^{-1} , and a slope parameter, $S(\tau)$, of 15.81 mm hr^{-1} for the ATI approach. Regression analysis for the HART method, involving precipitation efficiencies of Texas, resulted in $\tau = 5 \text{ mm hr}^{-1}$ and $S(\tau) = 11.86 \text{ mm hr}^{-1}$. These slopes and thresholds were determined using rainrate fields estimated from the DBZ samples, where the convective or stratiform ZR relation was applied to those reflectivity samples labeled convective or stratiform, respectively. The use of convective and stratiform ZR relations was prompted by the uncertainty expressed by Atlas et al. (1990) over the use of a single ZR relation for the measurement of all rainfall.

Also, choosing the best slope parameter based on correlation coefficients raises additional uncertainty in that the correlation coefficients for $4 \leq \tau \leq 6$ were so close that significant digits becomes an issue. Only with four significant figures does $\tau = 5$ become the best correlation coefficient. Since the climatological rainrate, $S(\tau)$, and echo area will vary with choice of τ so will the volume rain estimate. This suggests the need for additional calibration between volume estimates and measured surface rain volumes as a means for choosing the most accurate

climatological rainrate. To this end accurate recording of the spatial distribution of surface rainfall is critical, and considering the horizontal extent of individual convective elements, could be accomplished using a surface gauge network with 0.5 km horizontal resolution.

Results of volumetric precipitation estimates using a rainrate threshold of 5 mm hr^{-1} with each regression line combination are presented in Table 4, along with actual volumetric rainfall amounts obtained from surface rain-gauges. Actual daily volumetric totals were obtained by first mapping the irregularly spaced daily raingauge amounts into a $720 \times 720 \text{ km}^2$ rectangular grid centered on the storms which provided the radar data. This was performed in the same manner as that for the low resolution infrared data. After the mapping, actual daily rain volumes were simply computed as the sum of rain volumes at each grid point within the analysis domain. Due to the scarcity of the Mexican rainfall data and the highly variable spatial distribution of surface rainfall, the mapping of the irregularly spaced gauge data into a regular grid introduces some uncertainty into the actual rain volumes computed.

Consideration of the precipitation estimates in terms of the regression equations used, indicates that the case specific regression lines produced the overall best estimates of surface rain volumes. Qualitatively, the better performance of the case specific equations can be seen using the

average ratio of estimated to actual precipitation and the standard deviation of this ratio. Since each regression equation pair produced nine rainfall estimates, these ratios were computed by comparing each of the nine rainfall estimates with the actual daily rain volume for each sample. For the case specific regression lines, the average ratio of estimated to observed was 1.32 with a standard deviation of 0.59. These numbers are much lower than the average ratios(3.77, 3.15) and standard deviations(3.06, 2.74) calculated using estimates from the best convective(17a) and best stratiform(12b) equations, and those resulting from combination of the two best convective and two best stratiform equations, respectively. This wasn't surprising in light of the correlation analysis.

In assessing the effectiveness of the grid point, ATI, and HART prediction schemes, analysis of the ratio of estimated to actual rain volumes, and the standard deviation of these ratios, were again considered. In this case, the average ratio and standard deviations of each approach were determined by comparing all such ratios within each of the three methods. Overall, the grid point method had the lowest average ratio(2.34) and standard deviation of such ratios(2.23), while the ATI approach showed the highest in both cases, displaying an average ratio of 3.37 and standard deviation of 3.11. The HART method fell between these

two, and very close to the grid point scheme, with an average ratio of 2.53 and standard deviation of 2.35.

Comparison of the HART and the ATI approaches, indicated that the ATI estimates were higher than the HART estimates by a factor of ~ 1.33 in all cases. This made sense as the two methods differed only by their slope parameters, where the ATI slope parameter was ~ 1.33 times larger than that for HART. Overall, the HART method produced much better volume estimates than the ATI method. Since both methods are area-time integral approaches, the better estimates by HART were a direct result of including precipitation efficiencies. These had the effect of reducing the slope parameter for HART, and therefore, the precipitation estimates.

Unlike the ATI and HART approaches, the relative performances of the grid point and HART methods were so similar that, due to the uncertainty surrounding the actual daily rain volumes, it is hard to say which was better. Furthermore, considering the HART estimates obtained using $S(\tau=4 \text{ mm hr}^{-1})$, whose correlation coefficient was approximately equal to that for $S(\tau=5 \text{ mm hr}^{-1})$, we find that HART now displays the lowest average ratio of estimated to actual rain volumes (2.18) and the lowest standard deviation of such ratio (1.91). Clearly, any conclusion regarding the relative effectiveness of these two schemes will depend on the choice of $S(\tau)$. This further emphasizes the need for

accurate surface measurements.

Additional uncertainty arises through consideration of the slope parameters computed for the area-time integral approaches. In general, the HART and ATI methods have been found to be effective schemes for estimating precipitation from either ground based or space borne radars, provided that one obtains a representative radar sample of convective cells in varying stages of development. If this condition is met then, $S(\tau)$, obtained from the precipitation probability density function(pdf) using (4), is approximately equal to the slope parameter obtained through a linear regression of $\langle R \rangle$ and $F(\tau_i)$ pairs (Rosenfeld et al. 1990; Atlas et al. 1990).

In this experiment it remained to be seen if the nine random radar samples taken from three different convective storms provided a representative sample of convective cells. Comparison of $S(\tau)$ obtained through the regression of areawide rainrates against fractional coverages without consideration of precipitation efficiencies was 15.81 mm hr^{-1} , while $S(\tau)$ obtained using (4) was 20.23 mm hr^{-1} . This difference in the slope parameters suggests that the rainfall pdf obtained from the random radar samples was inadequate, or in other words, they failed to provide a distribution of rainrates representative of the true population.

The relative inaccuracy of $S(\tau)$ obtained from the pdf is further indicated by the effective slope parameters, or

$S(\tau)$ computed from the areal distribution of estimated radar reflectivities and listed in column five of Table 4. All but one of these parameters tend toward the lower slope parameter obtained from the regression analysis of average areawide rainrates and fractional coverages. In consideration of this uncertainty, the relative accuracy and utility of each of the three precipitation estimation schemes utilized in this analysis remains a point of conjecture.

7. Summary and Conclusions

The preceding analysis implemented and evaluated a new satellite rainfall estimation technique. First, how digital IR counts were related to vertically averaged internal radar reflectivities was determined. After positive linear relationships were found, the possibility of a climatological, convective or stratiform, linear DBZ-IR relation was examined. Finally, the linear relations derived in phase one were used to estimate daily surface rainfall.

The regression analysis of part one showed that a significant positive relationship between digital IR counts and vertically averaged internal radar reflectivities existed for most stratiform samples and for convective samples of MCS's in the early stages of their lifecycle. The greater consistency shown among stratiform types was felt to result from the greater homogeneity of both the IR and DBZ fields associated with these clouds. Fewer significant

correlations were found among the convective samples due to the highly cellular and dynamical nature of these regions and the obscuring effects of anvil development.

Finding positive correlations in part one of this analysis lead to the investigation of the nature of these relationships. Were these relations dictated by the system type and climatology of the day? Or did universal relationships exist among stratiform, convective or all types? All attempts to combine regression lines, both within or across cases, met with failure. Consequently, for the data used in this analysis, it was concluded that any linear relations between digital IR counts and internal radar reflectivities were tenuous, and varied with cloud type, stage of development, and climatology of the day.

The final aspect of this study involved estimating daily surface rain volumes from GOES images using three rainfall estimation techniques in tandem with three combinations of regression equations. In general, both the HART and grid point approach out performed the ATI method. Definitive conclusions could not be drawn, however, due to the uncertainties surrounding the actual surface rain volumes, the choice of $S(\tau)$, and the suggested misrepresentation of the precipitation probability density function by the nine vertically averaged horizontal radar reflectivity samples.

In terms of regression equation pairings, it was found that storm specific application of these equations per-

formed more consistently than the other two combinations, which resulted in precipitation estimates as much as eight times larger than the approximate daily rain volumes. The case specific equations were typically within a factor of two.

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APPENDIX

Calculating Degrees of Freedom

The assignment of degrees of freedom for any statistical study depends on the number of independent pieces of information available for the decision making process. Typically, one assumes that all values entering into a correlation or regression analysis are independent of each other. This was not the case in this study.

First of all, there was the problem of different resolutions between the two data sets. With the horizontal resolution of the infrared and reflectivity grids being, 8 km and 3 km, respectively, there was a different number of points representing the same grid area. As a result, the number of degrees of freedom for each image would not be the same.

Secondly, both images displayed some degree of spatial correlation among their component grid points. In this case, the IR or DBZ value at a particular grid point was not independent of its neighbors and therefore, the number of degrees of for each image could not be equal to the number of points within that image.

To address these problems, degrees of freedom were treated as a function of the number of non-zero grid points in each grid and the number of spatially correlated IR and DBZ values, respectively. Accordingly, degrees of freedom

were estimated by dividing the number of non-zero grid points used in the regression analysis by the average number of grid points surrounding any particular point and showing a correlation of 50% or greater. This was performed separately on the vertically averaged horizontal reflectivity grids and the corresponding optimum infrared shields listed in Table 1. The optimized infrared images were used as a means of eliminating the differing IR and DBZ spatial resolutions from the degrees of freedom estimation.

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Figure 1.

Vertically averaged horizontal radar reflectivities and corresponding optimal infrared images for two samples listed in Table 1. Each tick mark represents 3 km. Units are DBZ and Digital IR counts for contoured reflectivities and digital IR counts, respectively.

Figure 2.

Infrared subsectors extracted from full sector GOES images.

Figure 3.

Scatter diagrams for each IR-DBZ pair used in the analysis. Horizontal axes are digital IR counts. Vertical axes are DBZ.

Table 1.

Cases Studied and Sample Identifiers. List of internal radar reflectivity samples analyzed. Day of storm, cloud type, duration of radar observation, and sample identifiers are included.

Table 2.

Correlation and Regression Results. Includes type of cloud sampled, C(convective), or S(stratiform), correlation coefficient(r), 95% confidence interval(CI) for the correlation coefficient, slope of regression line(b), range of degrees of freedom(from IR to DBZ values), IR and DBZ covariance, standard deviation of IR and DBZ values within individual IR and DBZ fields of each case, and the regression variance(s^2). All entries are organized in descending order by their respective correlation coefficients.

Table 3.

Regression Equations for each DBZ and IR pair.

Table 4.

Rainfall Estimation Results. Summary of daily rain volume estimation for each case study using ATI, HART, and grid point summations. Units for Slopes are mm hr^{-1} and volumes are mm^3 .

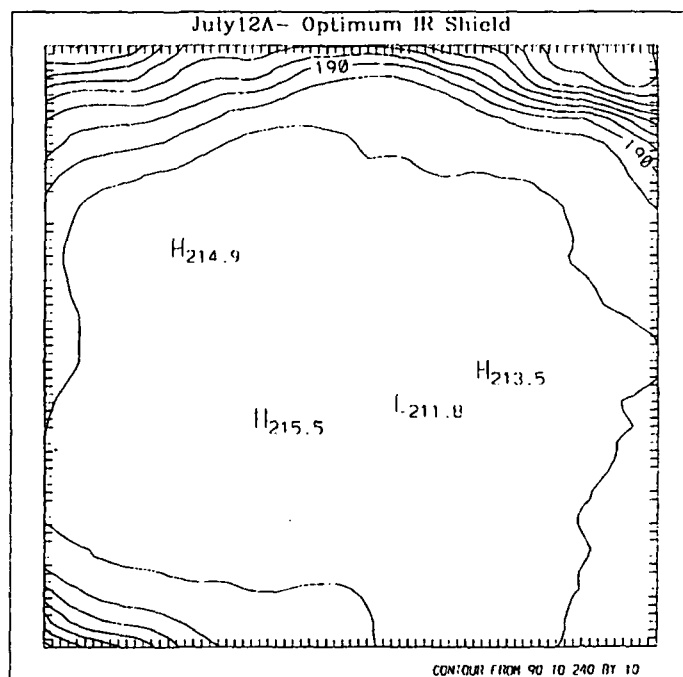
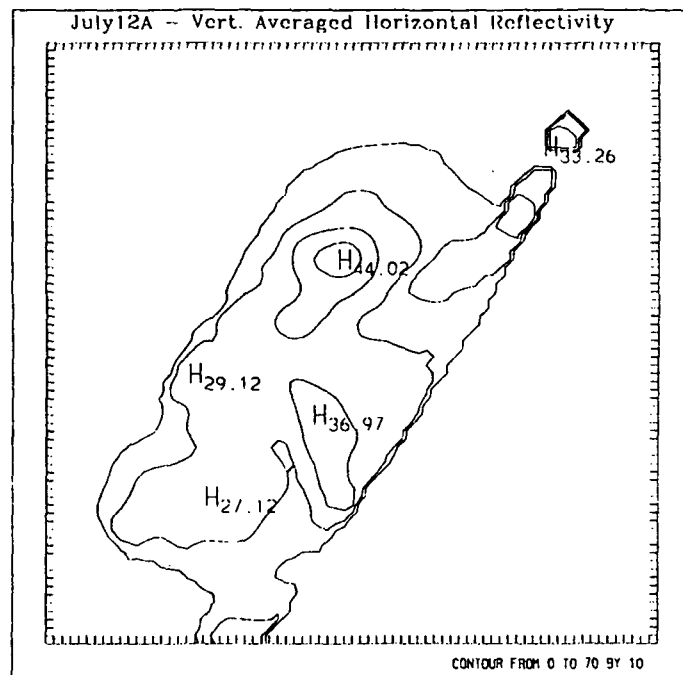


Figure 1. Vert. avgd. horizontal radar reflectivities and corresponding optimal infrared images for two samples listed in Table 1. Each tick mark represents 3 km. Units are DBZ and Digital IR counts.

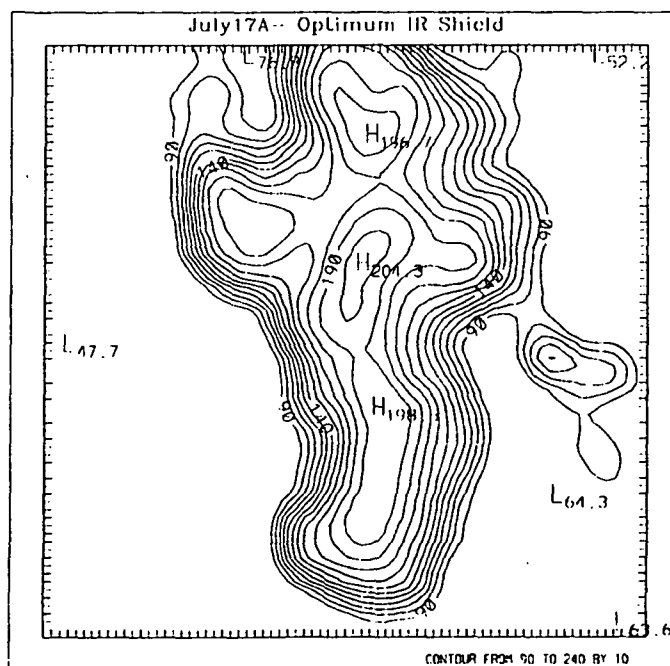
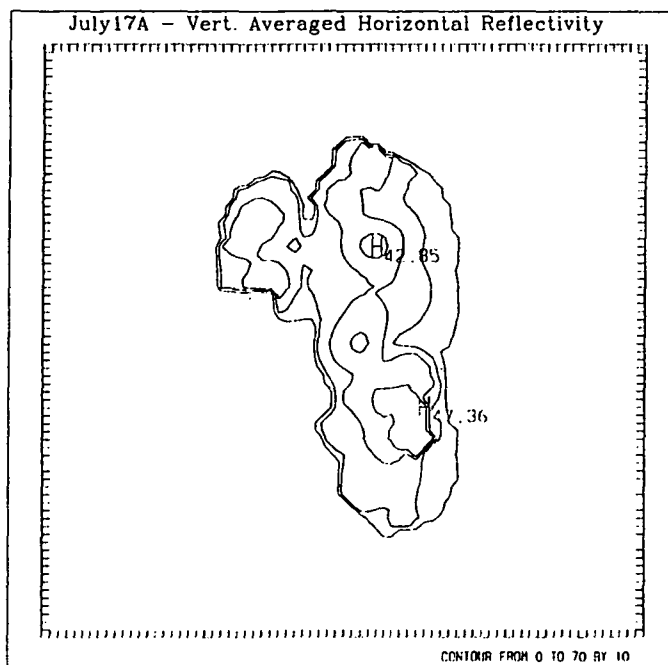
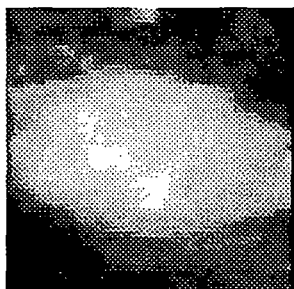
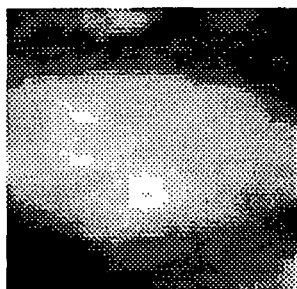


Figure 1. cont.

July12A(0334 UTC)



July12B(0438 UTC)



July12C(0453 UTC)



July17A(2043 UTC)



July17B(2112 UTC)



July17C(2247 UTC)



July20A(0250 UTC)



July20B(0310 UTC)



July20(0428 UTC)



Figure 2.

Infrared subsectors extracted from full sector GOES images.

<u>Case</u>	<u>Storm Type</u>	<u>Time(UTC)</u>	<u>Cloud Type</u>	<u>Sample ID</u>
12 July 1990	Meso-alpha	03:34-03:44	convective	July12A
		04:33-04:43	stratiform	July12B
		04:50-04:56	stratiform	July12C
17 July 1990	Multicell	20:36-20:50	convective	July17A
		21:08-21:15	convective	July17B
		22:44-22:51	stratiform	July17C
20 July 1990	Meso-beta	02:43-02:57	convective	July20A
		03:00-03:20	convective	July20B
		04:20-04:36	stratiform	July20C

Table 1. Cases Studied and Sample Identifiers

List of internal reflectivity samples analyzed. Day of storm, cloud type, duration of radar observation, and sample identifiers are also included.

<u>Sample</u>	<u>Type</u>	<u>Corr.</u> <u>Coef(r)</u>	<u>95%</u> <u>CI</u>	<u>Slope</u> <u>(b)</u>	<u>Degrees</u> <u>Freedom</u>	<u>Covar</u>	<u>Std. Dev.</u>		<u>Reg.</u> <u>Var.(s²)</u>
							<u>DBZ</u>	<u>IR</u>	
July12B	S	0.815	.80-.84	2.47	4-14	10.04	6.11	2.02	12.52
July12C	S	0.708	.68-.74	1.38	5-17	21.36	7.67	3.93	29.38
July20C	S	0.600	.57-.63	0.80	6-22	30.30	8.20	6.16	42.97
July17A	C	0.584	.54-.62	0.26	3-12	108.35	9.14	20.31	55.05
July17B	C	0.576	.53-.63	0.24	2- 8	95.97	8.37	19.93	46.73
July20A	C	0.531	.49-.57	0.19	4-14	183.01	11.04	31.22	87.60
July12A	C	0.507	.48-.54	1.64	7-24	11.49	8.56	2.65	54.51
July20B	C	0.339	.29-.39	0.01	4-16	125.71	10.43	35.53	96.33
July17C	S	0.314	.25-.37	0.25	3-12	29.53	8.67	10.85	67.79

Table 2.

Correlation and Regression Results. Includes type of cloud sampled, C(convective), or S(stratiform), correlation coefficient(r), 95% confidence interval(CI) for the correlation coefficient, slope of regression line(b), range of degrees of freedom(from IR to DBZ values), IR and DBZ covariance, standard deviation of IR and DBZ values within individual IR and DBZ fields of each case, and the regression variance(s²). All entries are organized in descending order by their respective correlation coefficients.

<u>Sample</u>	<u>Eqn.No.</u>	<u>Least Squares Regression Lines</u>
July12A	12a	$DBZ=1.64(IR-212.02)+19.74$
July12B	12b	$DBZ=2.47(IR-211.07)+22.05$
July12C	12c	$DBZ=1.38(IR-208.05)+20.21$
July17A	17a	$DBZ=0.26(IR-174.91)+25.81$
July17B	17b	$DBZ=0.24(IR-189.70)+24.56$
July17C	17c	$DBZ=0.25(IR-194.50)+18.67$
July20A	20a	$DBZ=0.18(IR-178.08)+24.09$
July20B	20b	$DBZ=0.10(IR-173.29)+22.87$
July20C	20c	$DBZ=0.80(IR-206.83)+18.26$

Table 3.

Regression Equations for each DBZ and IR pair.

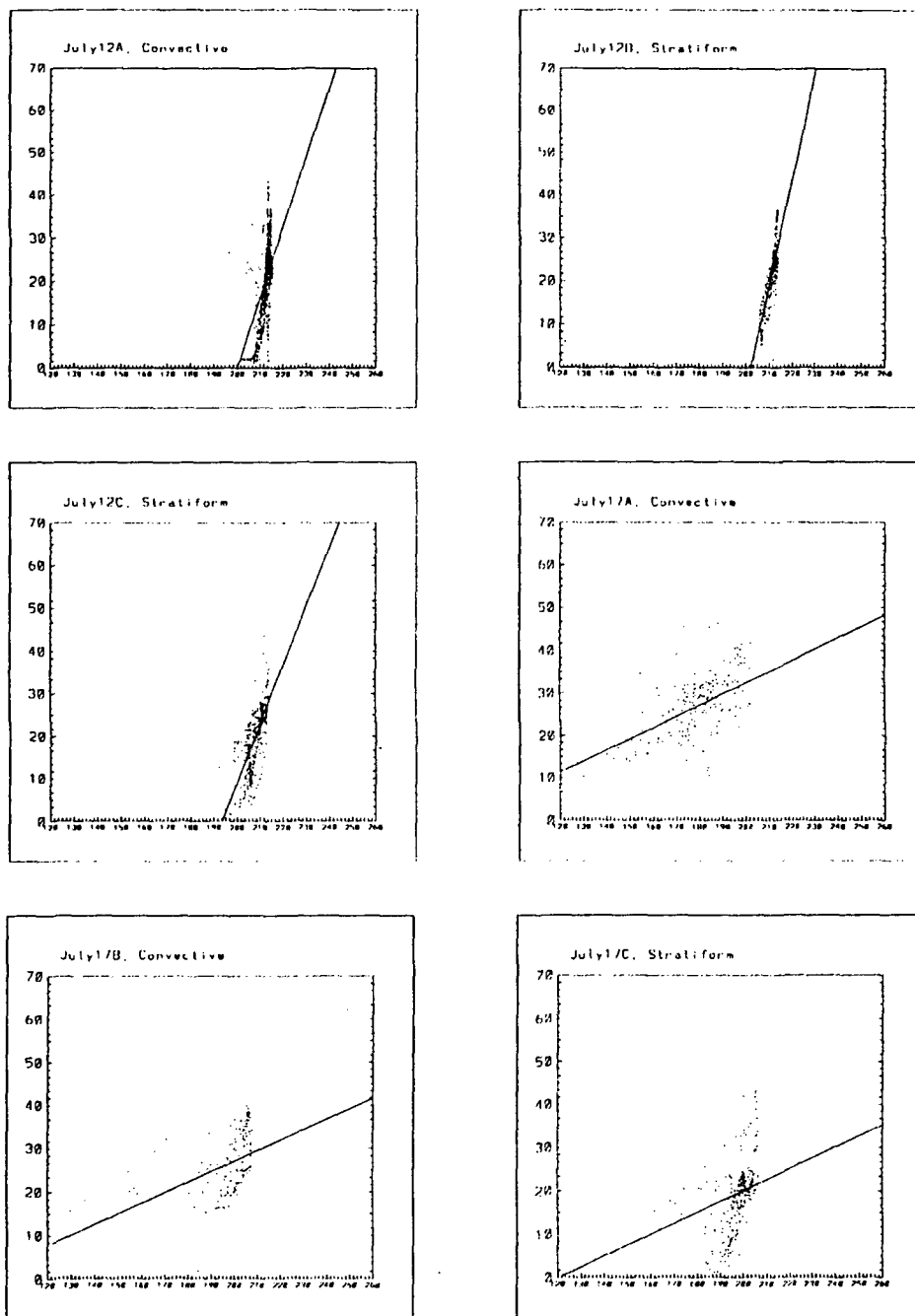


Figure 3.

Scatter diagrams for each IR-DBZ pair used in the analysis. Horizontal axes are digital IR counts. Vertical axes are DBZ.

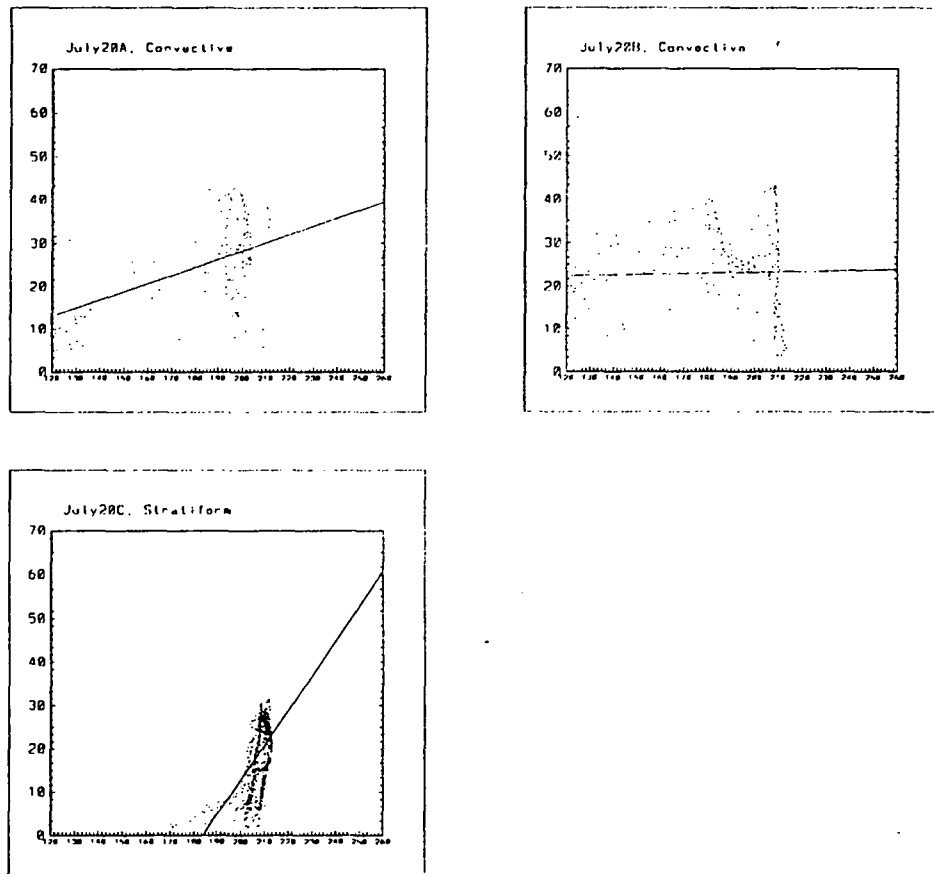


Figure 3 - Scatter Diagrams (cont)

a. Best Convective and Stratiform Regression Equations for each case.

	<u>ATI Method</u>		<u>HART Method</u>		<u>Grid Point Method</u>	
	<u>slope</u>	<u>volume</u>	<u>slope</u>	<u>volume</u>	<u>Slope</u>	<u>volume</u>
12 July 1990	15.81	2.62×10^{18}	11.86	1.59×10^{18}	22.69	3.04×10^{18}
17 July 1990	15.81	7.00×10^{18}	11.86	5.25×10^{18}	12.04	5.33×10^{18}
20 July 1990	15.81	8.19×10^{18}	11.86	6.14×10^{18}	8.29	4.30×10^{18}

b. Best Overall Convective and Stratiform Regression Equations

	<u>ATI Method</u>		<u>HART Method</u>		<u>Grid Point Method</u>	
	<u>slope</u>	<u>volume</u>	<u>slope</u>	<u>volume</u>	<u>Slope</u>	<u>volume</u>
12 July 1990	15.81	2.19×10^{19}	11.86	1.65×10^{19}	12.93	1.79×10^{19}
17 July 1990	15.81	7.00×10^{18}	11.86	5.25×10^{18}	12.04	5.33×10^{18}
20 July 1990	15.81	1.14×10^{19}	11.86	8.56×10^{18}	11.54	8.33×10^{18}

c. Combination of Two Best Convective and Two Best Stratiform Regression Lines

	<u>ATI Method</u>		<u>HART Method</u>		<u>Grid Point Method</u>	
	<u>slope</u>	<u>volume</u>	<u>slope</u>	<u>volume</u>	<u>Slope</u>	<u>volume</u>
12 July 1990	15.81	2.06×10^{19}	11.86	1.55×10^{19}	8.65	1.13×10^{19}
17 July 1990	15.81	6.40×10^{18}	11.86	4.80×10^{18}	8.32	3.37×10^{18}
20 July 1990	15.81	9.97×10^{18}	11.86	7.48×10^{18}	8.63	5.44×10^{18}

d. Actual Rainfall Volumes

12 July 1990	2.42×10^{18}
17 July 1990	2.91×10^{18}
20 July 1990	6.11×10^{18}

Table 4.

Rainfall Estimation Results. Summary of daily rain volume estimation for each case study using ATI, HART, and grid point summations. Units for Slopes are mm hr^{-1} and volumes are mm^3 .

APPENDIX II — Space-Time Variability of the Mexican Monsoon

Space-Time Variability of the Mexican Monsoon

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abstract

Rainfall from the Mexican monsoon is difficult to measure or interpret because this rainfall is particularly variable in time and space. Monthly and hourly long-term rain gage records were used to quantify rainfall variability in terms of the diurnal cycle, frequency distribution of monthly rainfall, and the spatial persistence of monthly rainfall anomalies. The diurnal cycle and form of the monthly frequency distribution were found to vary spatially within the area affected by the Mexican monsoon. Gage data were also used to construct a 30-year sequence of hourly precipitation averaged over a 4 degree by 5 degree grid (and percent area receiving rain). These areal data were used to compare point versus areal characteristics of rainfall and to provide the basis for satellite sampling experiments. The areal rainfall was sampled once every 13 hours in order to estimate temporal sampling errors of rainfall measured with the TRMM satellite. Three years of "satellite" sampling were found to be insufficient for identifying the diurnal cycle. The root mean square error of "satellite" estimates of August rainfall was 38%.

1. INTRODUCTION

Rainfall is an intermittent process that varies in time and space and can be considered a nonstationary stochastic processes. Characterization of this process depends on the time and space scales of interest. At a time scale of hours and a space scale of tens of kilometers, convective rainfall is considered to be highly variable. In terms of interannual variability, the largest variability, relative to mean conditions, tends to be found in regions of low precipitation. The Mexican Monsoon, which brings summer convection to the subtropical deserts of northwestern Mexico and the southwestern United States, is a particularly variable precipitation regime.

There are a number of circumstances when investigators working with rainfall data would benefit from a greater understanding of space-time variability of the Mexican monsoon. Often these circumstances revolve around some type of sampling problem. For example, rainfall is traditionally measured continuously at a few points and inferences are made about rainfall characteristics in other, unsampled locations. An unusual but interesting example involves paleoclimate reconstructions based on tree rings. How far from the tree ring site can the paleoclimate reconstructions be considered valid? A

different type of problem arises with satellites that measure rainfall continuously in space but discontinuously in time. The presence of a diurnal cycle can introduce biases into observations made under some types of satellite sampling schemes.

The goal of this paper is to characterize the variability of the rainfall from the Mexican Monsoon region, and to use this information to address some sampling and modeling problems. Where possible the study encompasses the entire Mexican Monsoon region (Figure 1), but because of data availability, many of the analyses are conducted only for the U.S. portion of the Monsoon.

Beyond a mere description of the long-term mean, there are many interesting aspects of the rainfall process:

- (1) the diurnal and seasonal cycles;
- (2) interannual variability;
- (3) intermittency and persistence;
- (4) extremes;
- (5) spatial variations in any of the above;
- (6) scaling between points and areas and other aspects of space-time organization of rainfall;
- (7) physics of entrainment, transport, and release of atmospheric moisture;
- (8) long-term climate change; and
- (9) relationship of rainfall characteristics to land surface and sea surface properties.

We cannot and do not address all these issues here, focusing instead on the diurnal cycle, intermittency, spatial variations in interannual variability and the diurnal cycle, the relationship between point and areal rain, as well as sampling problems that will be faced by satellite such as the Tropical Rainfall Measuring Mission (TRMM) satellite. Sampling problems were addressed using a 30 year sequence of hourly precipitation over a $4^{\circ} \times 5^{\circ}$ area. We wanted to address the following questions:

- 1) Given the sampling uncertainties, will three years of TRMM measurements be enough to identify the diurnal cycle?
- 2) Given the sampling uncertainties, how well will TRMM measure mean monthly rain over a three year period?

2. BACKGROUND

2.1 Mexican Monsoon

The Mexican monsoon begins in June in northwestern Mexico and is strongest during July, August and September. Rain can be organized as small localized

thunderstorms, tropical squall lines, or mesoscale systems (Smith and Gall, 1989). The onset is associated with high temperatures and wind shifts. It was long thought that moisture for the U.S. portion of the monsoon area came from the Gulf of Mexico. Current evidence, however, strongly suggests the moisture source is either the Gulf of California or the tropical eastern Pacific (Douglas et al., 1993; Hales, 1974).

2.2 Sampling problems

2.3 Modeling problems

Precipitation models may be divided into several classes: general circulation models of atmospheric processes including precipitation (applied globally or at the mesoscale), general circulation models used for short-term weather prediction, stochastic models of rainfall at a point, stochastic models of two dimensional rainfall fields, and regional models of precipitation dynamics (Barros and Lettenmaier, 1995). The stochastic models usually consider the occurrence of rain and the amount of rain (conditioned on rainfall occurrence) as separate processes (see Waymire and Gupta for a review). Simulation of stochastic processes normally entails random sampling of probability distribution functions representing some aspect of rainfall such as storm amount, storm duration, or number and size of convective cells. More attention has been devoted to the development of stochastic models than to the characterization of the underlying distribution functions, whose form and moments are often assumed rather than fit to observed data.

Increasing attention is being given to the general simulation of atmospheric processes because, among other factors, of the potential for relating precipitation to other climate variables. Unfortunately, precipitation is the most poorly simulated of the atmospheric processes (Morril, 1994; Mingchen and Dickinson). The best results can be expected with mesoscale models that are embedded within global models. Such a model has been developed for the Mexican Monsoon area (ref). Verification of GCM precipitation is difficult because of the natural variability of precipitation and limitations of precipitation data. Because GCM precipitation represents an areal average, it is necessary to compare simulated precipitation to observed values representing areal averages.

This paper provides information which is relevant to two types of modeling problems. First, it provides basic data that can be used for validation of simulated precipitation from mesoscale atmospheric models. Second, it provides basic data that can be used as the basis of constructing stochastic precipitation models.

3. METHODS

3.1 Data

1) AVAILABILITY

Much of the area affected by the Mexican monsoon is sparsely populated and sparsely gauged, particularly in Mexico. Raingages do, however, represent the only long-term (30-100 years) source of climatological data. Doppler radar (WSR-88D) is currently being installed in most of the U.S. portion of the monsoon region (Klazura and Imy, 1993). Satellite-based rainfall data are available for specific periods (Arkin et al., 1983; Negri et al., 1993). From 1997-1999, the Tropical Rainfall Measuring Mission (TRMM) will sample rainfall over the monsoon region (Simpson et al., 1988); TRMM will be a particularly important source of data over sparsely-gauged portions of Mexico.

2) THIS STUDY

Rain gage data were used for this study because of the availability of long term records and because gauge data is more accurate than remotely sensed precipitation estimates. Monthly rainfall for 120 long-term stations were obtained from a quality-controlled database for southwestern U.S. (Young, 1992). Monthly data for several additional stations were obtained from NOAA climate data (NOAA, various dates). Monthly data from several long-term Mexican stations were obtained from

the Mexican Climatological Service.

Hourly precipitation data were obtained from the Earth Info data set. The original source of the data is first order observing stations maintained by the U.S. National Weather Service.

Gauge data were not corrected for systematic measurement errors. Legates and DeLiberty (1993) estimate that in the Southwest, gauges underestimate summer rainfall by about 5% or 2 mm/month.

3.2 Areal Rainfall

1) HOURLY

A 30-year time series (1950-1979) of hourly areal rainfall was constructed for the months of July, August, and September. The areal average represents average rain over a 4° x 5° area in central Arizona (31° to 36° N. latitude and 109° to 113° W. longitude).

Areal averages for each hour were constructed by taking a weighted average of hourly gage rainfall. Because of the large number of computations (averages were computed for over 60,000 hourly time periods), a simple averaging method was needed. Ordinary kriging was investigated, but rejected when it was discovered that the network of operational gages changed frequently, which would have necessitated recomputation of the kriging weights. Instead, a normal-weighted method was used:

$$P^A = \sum_{i=1}^{ngages} \frac{N^A}{N^i} * P^i$$

where

P^A =areal rain

P^i =rain at gage i

N^A =normal rain over area

N^i =normal rain at gage i

The fraction of the $4^\circ \times 5^\circ$ grid receiving rainfall, F^A was estimated as:

$$F^A = \sum_{i=1}^{ngages} \frac{1}{ngages}$$

Data from sixteen gages were used. There was considerable missing data (which could be due to gage malfunction or failure to record and archive observations) and the data set was screened to discriminate between zero values associated with no rainfall and zero values associated with missing data. Thus the number of gages used in the averaging process ($ngages$ in above equations) varied depending on the number of missing gages.

2) MONTHLY

For monthly data, areal rainfall over one-degree cells was obtained by spatially interpolating gage data using detrended block kriging. The interpolation method which we used was developed by Michaud et al. (in press) in order to avoid biases that could be introduced by uneven sampling in the presence of trends introduced by orographic forcing or other factors.

3.3 Simulated Satellite Sampling

The simulated satellite "observations" were obtained by sampling observed hourly data (averaged over a $4^\circ \times 5^\circ$ grid) at regular intervals representing satellite overpasses. The orbit of the TRMM satellite is such that any site is sampled once every 13 hours. To simulate TRMM sampling, the observed hourly time series was "sampled" once every 13 hours for three years (the expected duration of the mission). Additional sampling experiments were conducted using a duration of 30 years (the duration of the ground data set). Because of the focus on the Mexican monsoon, sampling was conducted only for July, August and September.

4. CHARACTERIZATION OF SPACE-TIME VARIABILITY

Below, characterization of the space-time variability of rainfall is addressed using a variety of techniques and two temporal resolutions. Monthly data from long-term rain gage stations in the southwestern United States are used as the basis of a geostatistical analysis and an examination of the spatial variability of rainfall anomalies (deviations from the long term mean). Additional data from Mexican stations are used to address spatial variability in the frequency distribution of monthly rainfall. Hourly rain gage data from Arizona and New Mexico are used to examine the spatial variability of the diurnal cycle, the fractional coverage of rainfall, and the frequency distribution of hourly rainfall. In some cases, comparisons are made between point distributions and areal distributions.

4.1 Literature Review

This section is intended to provide an overview of what is already known about the space-time distribution of rainfall from the Mexican monsoon. Figure 1 depicts the geographic variability of long-term mean rainfall; rainfall is greatest along the western slopes of coastal mountains blocking the flow of moisture from the Pacific to the central plateaus of Northern Mexico. Rainfall decreases northward and northern Arizona may be considered to be the northern limit of the Mexican monsoon (sometimes called the "Arizona monsoon" in Arizona).

Michaud et al. (in press) used observations from a consistent 60 period to describe the spatial variation of mean summer rainfall in New Mexico, Arizona, southern Nevada and southern California. They found that summer rainfall within this area increases linearly with elevation at rates that range from 0.0003 to 0.0014 mm of rain per meter of elevation per day. Duckstein and Fogel (date) have demonstrated that this increase is because higher elevations receive both more storms and more rain per storm. Using satellite rainfall estimates, _____ (ref) found that monsoon rainfall along the west side of the Sierra Madre increases up an elevation of _____ meters, and then decreases.

4.2 Geostatistics

Geostatistical analysis offers a classic method of describing spatial variability (Issaks and Srivastava, 1989). Variograms were developed from monthly and hourly data (July, August, and September only). Sixty years of data (1915-1974) from 91 stations from New Mexico, Arizona, southern Nevada and Southern California were used for estimation of the monthly covariance functions and 30 years of data at 16 stations in Arizona were used for the hourly covariograms. Figure 2 compares the sample covariance obtained with monthly data to the sample covariance obtained with hourly data.

The range of the variograms indicates the approximate distance beyond which there is no autocorrelation. Based on correlograms fitted to monthly data, the range was 666, 727, and 996 km for July, August, and September, respectively. The range was 150 km for hourly data (based on a correlogram fitted to hourly

data from July, August, and September. Thus, interpolation beyond these distances would be unwise.

At monthly time scales both the mean and variance are nonstationary. Plotting station variance against station mean showed that the variance of monthly rainfall increases as a nonlinear function of mean rainfall. The coefficient of variation (CV) ranges from 3.0 for the driest stations to 0.4 for the wettest stations. For most stations, CV is near 0.5.

4.3 Monthly Anomalies

The spatial variability of rainfall anomalies (deviation from long-term mean) was examined as follows. Monthly gage observations were interpolated to obtain averages over one-degree cells (see section 3.2.2). Long-term mean rainfall from the gridded climatology of Michaud et al. (in press) was then subtracted from these values. Contour maps of the resulting anomalies were then prepared (Figure 3). Contour maps showing the interpolation variance were also prepared (but are not shown). These were examined to evaluate the possibility that apparent spatial trends in rainfall anomalies were due to interpolation uncertainty arising from sparse sampling and/or inconsistent anomalies between closely-spaced gages. This exercise indicated that the major trends in rainfall anomalies shown in figure 3 are real rather than spurious, although details in certain spots are less certain.

Contour maps of rainfall anomalies were examined for eight months. In three of the eight month, rainfall anomalies were more or less consistently low over the region. However, it was more common for rainfall anomalies to vary significantly from one part of the Southwest to the next. For example, in Figure 3a, a positive anomaly in central New Mexico (as much as + 30 mm or about 80% above the mean) is a sharp contrast to the negative anomaly in central Arizona (as much as -30 mm or about 80% below the mean). The anomaly maps for the eight months are broadly consistent with the geostatistical analysis conducted with 60 years of data, which showed that monthly rainfall is completely uncorrelated beyond about 700-1000 km, and strongly correlated ($r^2 > 0.50$) only for distances less than about 250 km.

4.4 Frequency Distribution

1) MONTHLY POINT DISTRIBUTIONS

The frequency distribution of monthly rainfall was examined using long-term records at 24 stations in the southwestern United States (Almagordo, Indio, Yuma, El Paso, Carlsbad, Nogales, Payson, Phoenix, Flagstaff, Tucson, Taos, Socorro, Albuquerque, Grand Canyon and Bisbee) and northwestern Mexico (Ensanada, Muleage, Hermosillo, Navojoa, Cananea, Mazatlan, Gihhuaha, and Monterray). The location of these stations is indicated in figure 1. The period of record used for the United States Stations was 1948-90; for Mexican stations it was 02-77 (minus numerous missing years).

Empirical distribution functions were prepared for each station for July, August, and September. Examination of these distributions showed that 44% of the stations had a bell-shaped distribution, 28% had bimodal distributions, and 25% had exponentially-shaped distributions. As a rule, the stations receiving the last rainfall had exponential, rather than bell-shaped, distributions. ____ of the stations had at least one summer month without rain.

Several frequency distributions (Kappa, Gamma, Poisson, and Weibull) were fit to the data; in each case the gamma distribution provided the best fit according to the Chi Squared criterion. Table 1 gives parameters of the fitted gamma distributions.

The above results could be useful to those constructing stochastic precipitation models or to those wishing to make statistical inferences about observations from short periods of record.

2) POINT VERSUS AREAL DISTRIBUTIONS OF HOURLY AND MONTHLY VALUES (rewrite this section)

Empirical cumulative frequency distributions were prepared from the hourly time series representing areal averages over a 4° x 5° grid in southern Arizona (section 3.2.1). In order to compare point and areal distributions, the same analysis was conducted for hourly point rainfall at one of the gages (Tucson) within the grid (Figure 4b-4d).

Examination of areal rainfall and the differences between point and areal rainfall revealed the following (for hourly summer rainfall in southern Arizona):

- (1) Rain was observed 3% of the time in the Tucson gage, whereas 22% of the time there was rain in at least one of the ~16 gages within the 4° x 5° grid. (With more complete sampling, the 22% value would be expected to increase somewhat.)
- (2) When it was raining somewhere within the grid, 11% of the grid was rainy, and 89% was dry, average. These percentages fluctuated slightly with time of day. Considering both rainy and dry periods, 99.86% of the time the percentage of the grid receiving rain was less than 50%.
- (3) Examination of empirical frequency distributions (see Figure 4 for August distributions) showed that very small amounts of rainfall are more prevalent in the areal data, whereas very large amounts of rainfall are more prevalent in the point data. For example, 0.1% of the areal values exceed 3.5 mm, whereas 0.1% of the point values exceed 13 mm.

The point and areal data examined above were averaged to a monthly basis so that hourly distributions could be compared to monthly distributions (Figure 4a). Over 30 summers, all point monthly values and all areal monthly values are nonzero. The point and areal distributions are fairly similar up to the 95% cumulative frequency. While the point distribution has a long upper tail, the areal monthly distribution completely lacks a long upper tail.

Table 1. Differences in variability of point and areal values: (check numbers)

	point (Tucson)	areal
hourly standard deviation	0.89 mm	0.26
monthly standard deviation	40.8	25.2
minimum value		
maximum value		

4.5 Diurnal Cycle

4.6 Intermittency

5. SATELLITE SAMPLING ERRORS

5.1 Diurnal Cycle

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6. DISCUSSION

References

World Meteorological Organization (WMO), 1979. Climate Atlas of North and Central America. World Meteorological Organization.

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APPENDIX III — Spatial and Elevational Variations of Summer Rainfall in the Southwestern
United States

SPATIAL AND ELEVATIONAL VARIATIONS OF SUMMER RAINFALL
IN THE SOUTHWESTERN UNITED STATES

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Abstract

This study examines the spatial variability of mean monthly summer rainfall in the southwestern United States with special attention given to the effect of elevation. Rain gage data from a consistent 60-year period show that mean rainfall increases linearly with elevation within a local area. A simple model (rain = normalized rainfall as a function of latitude and longitude + elevation coefficient * elevation) explains a large part of the spatial variability of mean rainfall. The rainfall model (the MSWR model) and digital elevation data were used to produce a $1^\circ \times 1^\circ$ gridded rainfall climatology for July, August, and September. Regional rainfall estimated with this model is 9.3% higher than an estimate based on arithmetic averaging of gage data over $2^\circ \times 2^\circ$ areas. For individual $2^\circ \times 2^\circ$ cells, the difference between model rainfall and the arithmetic mean of gage rainfall ranged from -250% to +41%.

The MSWR model was used to remove orographic effects from regional rainfall fields. When rainfall is normalized to sea level, two rainfall maximums emerge: one in southcentral Arizona associated with the Mexican monsoon maximum and one in southeastern New Mexico associated with the Gulf of Mexico. Detrended block kriging (using the MSWR model as an estimate of the long-term trend) and monthly rain gage data were used to produce unbiased areal rainfall estimates which were compared to $1^\circ \times 1^\circ$ satellite-based rainfall estimates. On a month-by-month basis, there were large differences between the two estimates, although the comparison improved after temporal averaging.

1. Introduction

a. Statement of the Problem

It is well-known that rainfall accumulated over a sufficient period of time varies with elevation, and that orographic effects must be taken into account when using or interpreting rainfall data in mountainous terrain. This study was undertaken because of a desire to account for the effect of topography on rainfall from the Mexican monsoon, which brings summer thunderstorms to the mountainous deserts of northwestern Mexico and the southwestern United States. Understanding how topography affects rainfall will help us to better interpret the spatial variability of monsoon rainfall. Understanding how topography affects rainfall is also important when interpolating or disaggregating rainfall data. There are many occasions when an investigator may wish to interpolate rain gage data in order to estimate rainfall at locations where it has not been measured. In mountainous terrain, a credible interpolation procedure must take nonstationarity into account, which in this case refers to spatial trends in the expected value (or spatial covariance) of the rainfall process. Nonstationarity in long-term mean rainfall can be due to orographic enhancement or other factors, such as prevailing patterns in the transport and release of atmospheric moisture.

b. Objectives

The main objective of this study was to produce an estimate of the spatial variability of mean summer rainfall in the southwestern United States, with consideration of the variability caused by elevation. The geographic scope (New Mexico, Arizona, and portions of southern California and Nevada) was chosen to include the most densely gaged portion of the area which is affected (or potentially affected) by the Mexican monsoon. The months of July, August, and September were examined separately to gain insights into seasonality.

There were two motivations for quantifying rainfall variability. First, this provides a basis for distinguishing between rainfall patterns due to spatial variability in monsoon strength and

rainfall patterns due to the orographic enhancement of rainfall. Second, a description of rainfall variability can be used to produce unbiased, areal precipitation estimates using rain gage data and interpolation techniques which explicitly recognize expected spatial trends. The paper concludes by demonstrating the use of such interpolation techniques for obtaining gage-based areal rainfall estimates. The gage-based rainfall estimates are compared to satellite-based rainfall estimates at a monthly, $1^\circ \times 1^\circ$ resolution.

c. Literature Review

1) EFFECT OF TOPOGRAPHY ON RAINFALL

There are several mechanisms that can cause rainfall to increase with elevation: mechanical lifting (resulting in cooling and condensation) as air is forced over a mountain barrier, enhanced initiation of convection over rough terrain, reductions in the virga effect (evaporation of rain as it falls), the "seeder-feeder" mechanism, and topographic retardation of the movement of low pressure systems. The seeder-feeder theory (drops from clouds aloft washing out small drops within low-level clouds formed above hills) was proposed by Bergeron (1968) after observing enhancement of rainfall over 30-40 meter hills. Rainfall which has been initiated on the windward side of a topographic barrier may persist some distance downwind of the summit (Hanson, 1982; Johnson and Hanson, 1995); "rain shadows" on the lee side of the barrier may be associated with atmospheric subsidence (Sumner, 1988). Moisture availability, exposure to mean flow, atmospheric stability, and topographic funneling of storms along preferred tracks are additional factors which can impact rainfall variability.

It is likely that reductions in the virga effect and enhanced initiation of convection over rough terrain are of greater importance for summer rainfall in the Southwest than for rainfall in other areas. Conversely, mechanisms such as mechanical lifting are probably of greater importance for frontal rainfall in the Northwest than for convective rainfall in the Southwest. It

is possible that regional variability in orographic mechanisms may affect the regional nature of the relationship between precipitation and elevation.

The effect of topography on rainfall is usually determined empirically on a regional basis, even though it is also possible to model this relationship physically (see, for example, Barros and Lettenmaier, 1993, 1994). While some of the empirical studies have examined data from individual storms, it is more common to work with long-term means in order to reduce random variability. It has been found that the topographic effect can vary seasonally (Karneili and Osborn, 1988; Hanson and Johnson, 1993). Most studies have found that gage rainfall is a linear or nonlinear function of station elevation. Daly et al. (1994), however, found that mean precipitation rainfall in the Columbia Gorge of the Pacific Northwest is insensitive to station elevation. Instead, rainfall increases with the average elevation within a 5-minute grid enclosing the station. There are also cases when elevation alone does not explain rainfall variability. The classic studies of Sreen (1947) and Schermerhorn (1967), conducted in Oregon, Washington, and Colorado, found that rise, orientation, latitude, and barrier effects were also important. Working in a small mountain watershed in Idaho, Hanson and Johnson (1993) found that the seasonal distribution of precipitation was quite different for rain-dominated areas at low elevation and snow-dominated areas at high elevations. Hanson and Johnson (1993) also found that frequency-depth-duration characteristics vary with elevation. For durations greater than one hour, rainfall at a given recurrence interval increased with elevation. For durations less than one hour, however, rainfall at a given recurrence interval decreased with elevation.

Duckstein et al. (1973) examined 3-7 years of summer data collected at various elevations in a small steep mountain range (Santa Catalinas) in southern Arizona. They found that the number of storms per summer increases considerably with elevation (10.2 storms per 1000 meters), and that the mean rain per storm increases moderately with elevation (0.00097 mm/m). The increase in the number of storms with elevation presents a challenge when interpolating storm rainfall from low to high elevations.

Martinez-Goytre et al. (1994) found paleoflood evidence that watersheds on the south side of the Santa Catalina Mountains (facing the prevailing storm tracks) experience slightly larger floods than watersheds on the north side of the mountains.

Karneili and Osborn (1988) examined precipitation-elevation relationships for the state of Arizona using 158 rain gage stations with at least 30 years of record. They found a linear relationship between summer/winter rainfall and elevation when the state was divided into three regions. The regressions for the three regions explained 94%, 74%, and 67%, respectively, of the spatial variance of summer rainfall.

2) THE MEXICAN MONSOON

The thunderstorms that bring summer rain to the Southwest United States have been termed the "Arizona monsoon" or the "Southwest monsoon". While this monsoon supplies slightly more than half of the area's annual precipitation, its rainfall is characterized by high intermittency in time and space as well as large interannual variability. Recently, Douglas et al. (1993) proposed the name "Mexican monsoon" in recognition of the fact that the Arizona rain is merely the northern fringe of heavy summer rainfall which falls on the western slopes of the Sierra Madre Occidental in northwestern Mexico (Figure 1).

The Mexican monsoon begins in June in northwestern Mexico and is strongest during July, August, and September. Rain can be organized as small localized thunderstorms, tropical squall lines, or mesoscale systems (Smith and Gall, 1989). The onset is associated with high temperatures and wind shifts. It was long thought that moisture for the "Arizona monsoon" came from the Gulf of Mexico. Current evidence, however, suggests that the dominant moisture source is either the Gulf of California or the tropical eastern Pacific (Douglas et al., 1993; Hales, 1974), although it is likely that some high-level moisture comes from the Gulf of Mexico.

On the basis of satellite rainfall data, Negri et al. (1993) found that monsoon rainfall on the west side of the Sierra Madre Occidental increases with elevation up to 1000 m. From 1000 to 2500 m., however, rainfall decreases with elevation.

3) SPATIAL INTERPOLATION

There is an extensive body of literature on spatial interpolation of point (gage) observations of rainfall. This literature will not be reviewed except to note that considerable attention has been given to geostatistically based techniques such as kriging or optimal interpolation. Several comparison studies have favored the geostatistical techniques (Tabios and Salas, 1985; Creutin and Obled, 1982), although multiple discriminant analysis (Young, 1992) and a regression-based technique known as PRISM (Daly et al., 1994) have also performed well.

Several techniques have been applied to the interpolation of rain gage measurements under nonstationary or mountainous conditions (Chua and Bras, 1982; Daly et al., 1994; Hevesi et al., 1992; Dingman et al., 1988; Phillips et al., 1992; Tabios and Salas, 1985). These include (a) universal kriging with estimation of the trend within the kriging framework, (b) universal kriging with estimation of the trend outside of the kriging framework, (c) kriging of detrended data, (d) co-kriging with elevation as the auxiliary variable, and (e) PRISM. When using universal kriging, it is generally recommended that estimation of the trend be performed outside the kriging framework.

d. Data

1) AVAILABILITY

Much of the area affected by the Mexican monsoon is sparsely populated and sparsely gaged, particularly in Mexico. Rain gages do, however, represent the only long-term (30-100 year) source of climatological data. Doppler radar (WSR-88D) is currently being installed in most of the United States' portion of the monsoon region (Klazura and Imy, 1993). Satellite-

based rainfall data are available for specific periods (Arkin, 1983; Negri et al., 1993). From 1997-1999, the Tropical Rainfall Measuring Mission (TRMM) will be an important source of rainfall data, particularly over sparsely gaged portions of Mexico (Simpson et al., 1988).

2) THIS STUDY

Rain gage data were used for this study because of the availability of long-term records and because gage data are more accurate than remotely sensed precipitation estimates. Nearly all of the rain gage data used in this study were obtained from a long-term, quality-controlled database of monthly rainfall in the southwestern United States (Young, 1992). Data for several additional stations were obtained from NOAA climate data (NOAA, various years). Additional information on the processing of the rainfall data is given in Section 2a. Data were not corrected for systematic measurement errors. Legates and DeLiberty (1993) estimated that in the Southwest, gages underestimate summer rainfall by about 5% or 2 mm/month.

The satellite rainfall used in this study were developed by Negri et al. (1993). These data are based on twice daily (700 and 1900 LST) passive microwave (SSM/I) observations. A coupled cloud-radiative transfer model was used to relate 86-GHz brightness temperatures to rainfall intensity. The data have monthly, 0.15° resolution and are segregated into morning and evening values. In this study, the sum of morning and evening values was used in the satellite-gage comparison; morning values were used for quality control.

Elevation data used in this study were derived from ETOP5 digital elevation data. Five-minute data were averaged to obtain the average elevation over $1^\circ \times 1^\circ$ cells.

2. Spatial and Elevational Variability of Mean Rainfall (MSWR Model)

This section presents a model of the spatial and elevational variability of mean monthly rain in Arizona, New Mexico, and parts of southern California and Nevada. Hereafter, the model will be called the MSWR (mean Southwest rainfall) model. The model is intended to be

practical for routine applications. The structure of the model and the required input data were therefore kept as simple as possible. For example, the model does not require a Digital Elevation Model (DEM) for point applications. Neither does it require difficult-to-compute variables such as exposure and topographic barrier indices. Barrier indices were avoided partly because it is difficult to determine with confidence how much of the moisture comes from the Gulf of California/Eastern Pacific and how much comes from the Gulf of Mexico. Further, the direction(s) of moisture transport is likely to vary within the region and by season. Fortunately, the following model based merely on location and elevation explains a large portion of the variability of the observed data:

$$P_j(\text{latitude,longitude,elevation}) = B_j(\text{latitude,longitude}) + C_j(R^k) * \text{elevation} \quad (1)$$

where:

P_j = mean rainfall in month j ;

k = month (July, August, or September);

B_j = normalized rainfall parameter;

C_j = elevation coefficient; and

R^k = region of uniform elevation coefficient.

The elevation coefficient C represents the rate at which rainfall increases with elevation. Parameter B represents rainfall normalized to sea level. If rainfall within a local area is plotted against elevation and a line is drawn through the data, C is the slope of the line, and B is the intercept. Values of B and C vary with month and with location. B is patchwise variable with one-degree resolution (each $1^\circ \times 1^\circ$ cell has a different value). C is also patchwise variable, but there are only two "patches" for C . Patchwise variability of parameters means that the rainfall field estimated by the model (and model residuals) may contain discontinuities.

In terms of spatial geometry, the model describes a rainfall surface defined over the domain of the study area. The total rainfall surface is the sum of two other surfaces: one which describes the amount of rainfall that is associated with orographic enhancement ($C \cdot \text{elevation}$) and one which is associated nonorographic, or "normalized" rainfall (parameter B). The normalized rainfall surface is uniform within each $1^\circ \times 1^\circ$ cell, but varies from cell to cell. The surface associated with orographic rainfall is proportional to elevation, but the constant of proportionality (C) is different for the deserts of southern California and Nevada than for the remainder of the study area. When a Digital Elevation Model (DEM) is used to describe topography, the spatial resolution of the orographic rainfall surface is related to the resolution of the DEM.

The MSWR model can be evaluated to give monthly mean rainfall for any point of known latitude, longitude, and elevation; this is analogous to sampling the total rainfall surface at a point. The model can be used, for example, to predict rain at the location of a rain gage station. In this case, the value of B is taken from the $1^\circ \times 1^\circ$ cell in which the station is located; the elevation coefficient is applied to the station elevation. Later in this paper (during model fitting and testing), observed gage rainfall is compared to point model rainfall evaluated at the gage location.

The model can also be used to estimate rain over areas; this is analogous to spatially integrating the total rainfall surface to obtain an area-average. In the areal case, the value of B is the areal average of B , and the elevation coefficient is applied to the mean elevation of the area in question. In the kriging techniques described in Section 3b, the MSWR model is evaluated for both points (to determine long-term mean rainfall at each gage) and for areas (to determine long-term areal rainfall for the estimation cell).

a. Calibration and Validation Data

Monthly means from 91 rain gage stations were used to calibrate the MSWR model, that is, to evaluate model parameters B_j and C_j using the observed station precipitation and station

elevation. The location of the calibration stations is shown in Figure 2. The following factors were taken into account when selecting calibration stations: (1) length of record, (2) concurrent periods of record, and (3) adequacy of spatial sampling. The first two criteria are important because the spatial and interannual variability of rainfall is very high. In Tucson, Arizona, for example, 164 years of record are needed to estimate mean July rain within $\pm 10\%$ (at 95% confidence, assuming the mean is normally distributed with mean m_x and variance σ_x^2/n). Continuous gage records rarely go back more than 50-80 years; therefore, it is difficult to determine mean rainfall with high precision. The selected period of record is the longest possible, given the constraints of adequate spatial sampling.

The period 1915-1974 was used for computing monthly means. Eighty-two stations had at least 90% of summer data available for this period. Nine additional stations with shorter or less-consistent periods of record were added to fill in data-sparse regions. Altogether, only 3% of summer data were missing from the final data set; missing or estimated data were neglected when computing monthly means.

An additional 26 stations were used for model validation (Figure 2). These are stations from Young's database that have at least 30 years of record during 1915-1974 (data outside this period was excluded).

b. Model Fitting

Plots of mean monthly station rain versus station elevation (and mean summer rain versus elevation) show that elevation alone does not explain rainfall variability across the study area (Figure 3). It was found, however, that if the study area is divided into certain subregions, rain within each subregion increases linearly with station elevation. This phenomenon can be seen in Figure 3 by examining the plotting symbols, which indicate the subregion. Subregions were selected subjectively on the basis of data groups in the precipitation-elevation plots. In order of decreasing rainfall, the subregions are: southwestern Arizona, central mountains of Arizona and

New Mexico, Colorado Plateau, and the California and Nevada deserts. Two stations in extreme southeast New Mexico, however, have as much rain as stations in southwestern Arizona (at similar elevations).

The elevation coefficient C (mm of rain per meter of elevation) in (1) was evaluated by regressing rain against elevation for each month and each of the four subregions. The regression coefficient was taken as an estimate of C , and the intercept was taken as an estimate of B . The result of the four regressions was a rainfall model similar to (1), in which the value of B was uniform within each subregion, but varied from subregion to subregion. The C values obtained from the regressions were identical for three of the subregions. C values were quite low, however, in the California/Nevada subregion (west of the dotted line in Figure 2). This subregion also has anomalously low rainfall, and the elevation coefficient is the same for July, August, and September. In contrast, the rest of the study area (composed of the three other subregions) has the same elevation coefficient for July and August and a much smaller coefficient for September.

Although rainfall variability was fairly well-described by the four-subregion model, residuals from that model displayed spatial trends. These trends suggested that allowing B to vary more continuously in space would provide a better fit to the data. Hence, values of B were estimated for one-degree cells.

In a final step, (1) was fit to the data using an iterative fitting procedure based on regression (for the elevation coefficient) and inverse distance weighting (for the normalized rainfall parameter). The first step in each iteration was to produce a residual for each station which was equal to the observed rainfall minus C times the station elevation. These elevation-detrended residuals were then used to update values of B . The B value for a particular one-degree cell was based on inverse distance weighting of residuals for stations located within that cell or within adjacent cells. (Inverse distance weighting was used instead of kriging because of the complications of estimating the covariance structure of a variable which is estimated iteratively.) The second step was to calculate a new type of residual for each station which was

equal to observed station rainfall minus the B value for the one-degree cell containing the station. This new residual was then regressed against station elevation to obtain an updated elevation coefficient C . Regressions were performed separately for the "dry" region and the main part of the study area. The entire procedure was then repeated until convergence was obtained. The fitting procedure was performed separately for each month. A weakness of the above procedure is that it does not guarantee orthogonality between normalized rainfall and the elevation component.

c. Results

1) PARAMETER VALUES AND ACCURACY ASSESSMENT

The MSWR model of mean monthly rain is given by (1), along with values of the normalized rainfall parameter B (Table 1) and elevation coefficient C (Table 2). Values of B and C vary according to month; summer rainfall can be obtained by adding values for July, August, and September. In each month, there are two values of C : one which applies to the main part of the study region and one which applies to the "dry" region. The extent of the dry region (see Figure 2) was determined from station rainfall and the assumption that topographic ridges separated the dry region from the main part of the study area.

Scatterplots of predicted versus observed point rain at the calibration and validation stations are shown in Figure 4; statistics summarizing model accuracy are given in Table 3. The MSWR model explains 92% of the variance of the data used for parameter estimation and 70% of the variance of data from 26 independent validation stations. However, some of the reduction in r^2 values for the validation data may be due to shorter periods of record which result in poorer estimates of long-term means. For example, there is a tendency for validation stations whose periods of record include relatively more years that are regionally dry (and less that are regionally wet) to have observed means which are significantly less than model predictions. Another

limitation of the validation data is that there are significant portions of the study area that contain few validation stations.

2) GRIDDED CLIMATOLOGY

The MSWR model and elevation data (mean elevation over 1° x 1° cells) were used to generate a gridded rainfall climatology representing area-average rain over 1° x 1° cells (Table 4).

3) COMPARISON WITH A SIMPLE MODEL

The MSWR model was compared to a simpler model as a reference. The simple model has the form (after Briggs and Cogley, 1995):

$$P_j = a_j * \text{latitude} + b_j * \text{longitude} + c_j * \text{elevation} + d_j \quad (2)$$

where P_j is mean rainfall in month j . Coefficients a , b , c , and d were evaluated by least squares for the entire study area. The simple model explains 72% of the variance of calibration data versus 92% for the MSWR model. For the validation data, the simple model explains 64% of the data variance compared to 70% for the MSWR model.

4) LOW ELEVATION BIAS OF RAIN GAGES

Areal rainfall was estimated with and without consideration of the low-elevation bias of rain gage stations in order to evaluate the strength of this bias. The study area was divided into 21 2° x 2° cells. Mean monthly rainfall was estimated for each cell using (a) the MSWR model and elevation data averaged over the two-degree cells, and (b) the arithmetic average of rain from stations within the cell. Both estimates were derived from the same data. However, the model-based estimate takes into account the difference between station elevation and mean elevation

within the cell. The model-based estimate also takes into account spatial trends not related to topography. On the average, the model-based rainfall estimate was 9.3% higher than rainfall based on arithmetic averaging. Results were more variable for individual $2^\circ \times 2^\circ$ cells (Table 5). For individual two-degree cells, the difference between the two estimates ranged from -250% (gages severely underestimate rain) to +41% (gages overestimate rain). Some of the greatest discrepancies occurred in sparsely gaged areas.

The combination of low-elevation gage bias (9% for the region) and gage undercatch (about 5%; see Legates and DeLiberty, 1993) could be of modest significance in some hydroclimatological applications. Results for the $2^\circ \times 2^\circ$ cells, however, suggest that bias could be a potentially serious problem when gage rainfall is used for hydrologic or ecologic modeling at the scale of about 10^4 km^2 .

5) INTERANNUAL VARIABILITY OF THE ELEVATION COEFFICIENT

Data from individual years show that the elevation coefficient ($C = \delta\text{rain}/\delta\text{elevation}$) is not consistent from year to year and tends to be slightly larger in wet years. Consistency of the elevation coefficient was evaluated by regressing station rainfall for an individual month against station elevation to obtain an elevation coefficient for that month (Figure 5). Only stations within the lower deserts of southcentral Arizona were used because rainfall within this region varies mostly with elevation and little with latitude and longitude. Data from 60 Augusts (1915-1974) were used to obtain 60 elevation coefficients. The elevation coefficients thus obtained ranged from 0.015 to 0.111 mm m^{-1} and tended to be slightly larger in months with high rainfall and slightly smaller when rainfall was low. The elevation coefficient for 12 of the Augusts was significantly different than the climatological coefficient derived from long-term mean rainfall.

6) DISCUSSION

The MSWR model is quite simple and can only be expected to account for gross spatial variations in mean rainfall. The model is surprisingly accurate considering that it does not explicitly account for aspect in relation to mean windflow, rainshadows, and topographically induced preferred storm tracks. Consideration of these variables may very well lead to improved prediction.

Values of MSWR parameters may be sensitive to changes in grid size. Based on an exploratory analysis, it is likely that least squares estimation will produce elevation coefficients that become smaller and normalized rainfall that becomes larger as cell size increases, say from 1° to 5° . It is unlikely that gage densities in mountainous areas would be sufficient to support parameter estimation for cells much smaller than 1° .

The "dry region" west of approximately 115°W (left of the dotted line in Figure 2) exhibits rainfall characteristics that are markedly different than those in the rest of the study area. This region may be associated with greater atmospheric stability. In addition, the 900 mb data of Douglas et al. (1993) suggest that the southerly low-level flow of moist air from Mexico to the United States occurs east of approximately 115°W .

Within the study region, summer rainfall increases linearly with station elevation at a rate of about 0.0003-0.0014 mm of rain per meter of elevation per day ($\text{mm m}^{-1} \text{d}^{-1}$). The most prevalent rate of increase is $0.0014 \text{ mm m}^{-1} \text{d}^{-1}$. These values, which represent long-term means, are within the range of rates that have been established for other locations. Within the Southwest, previous studies have obtained rates of $0.0013 \text{ mm m}^{-1} \text{d}^{-1}$ for summer rainfall in steep Arizona terrain (Duckstein et al., 1973), $\sim 0.0003 \text{ mm m}^{-1} \text{d}^{-1}$ for annual rainfall in southern Nevada (Hevesi et al., 1992; French, 1983), and $0.0034 \text{ mm m}^{-1} \text{d}^{-1}$ for winter rain in southwestern Colorado (Chua and Bras, 1982). The largest rate that we are aware of is $0.0114 \text{ mm m}^{-1} \text{d}^{-1}$ for annual rain in the Columbia Gorge in the Pacific Northwest (Daly et al., 1994).

Examination of data from individual years suggests that orographic enhancement of rainfall is slightly more pronounced in wet years than in dry years. This is consistent with the observation that high rates of orographic enhancement occur in the Pacific Northwest and that very low rates occur in the dry deserts of southern Nevada.

3. Applications of the MSWR Model

a. Spatial Patterns of Rainfall Normalized to Sea Level

Figure 6 depicts mean summer rainfall with and without the effect of elevation. The upper map gives spatial variations in total rainfall based on the MSWR rainfall climatology in Table 4; the lower map gives spatial variations in the MSWR model's normalized rainfall parameter (from Table 1). The lower map can be interpreted as rainfall normalized to sea level because the normalized rainfall parameter is the intercept of the regression line between rainfall and elevation.

The regional patterns of moisture transport and depletion are much more evident when rainfall is normalized to sea level (Figure 6b):

- (1) The northern tip of the Mexican monsoon maximum (which runs along the western flank of the coastal mountains in northwestern Mexico) can be seen in southcentral Arizona. While the rainfall maximum within Arizona is most pronounced at the Mexican border, the stippled area south of the 30 mm contour appears to be strongly associated with Mexican monsoon moisture, which presumably originates from the Gulf of California or tropical eastern Pacific. Topographically, the area within the 30 mm contour is a valley opening southwest to the Gulf of California.
- (2) A weaker rainfall maximum can be seen in southeastern New Mexico. This maximum, which is nearer to the Gulf of Mexico than any other portion of the study area, is presumably associated with moisture from the Gulf of Mexico. Normalized rainfall

decreases as one travels away from the Gulf of Mexico towards the mountains and high plateaus of northwestern New Mexico.

- (3) A rainfall "trough" running north-northwest in western New Mexico may be interpreted as the line dividing the region most strongly associated with Mexican monsoon moisture from the Gulf of California/tropical eastern Pacific and the region most strongly associated with moisture from the Gulf of Mexico. The exact location of this trough may be influenced by atmospheric stability over the north-south trending Rio Grande Valley.
- (4) Strongly negative values of normalized rainfall delineate the high deserts of northern Arizona and northwestern New Mexico.

Monthly maps similar to those in Figure 6b show that the rainfall maximum in eastern New Mexico is greatest in September, whereas the rainfall maximum along the Arizona/Mexico border is strongest in July. This seasonality is consistent with the seasonality noted by Douglas et al. (1993). In central Arizona, however, rainfall is greatest in August.

*b. Estimation of Areal Rainfall from Gage Data and Comparison with
Satellite-Based Rainfall Estimates*

Proper comparison of rain gage data and satellite rainfall estimates is hampered by the spatial incompatibility of point and areal measurements. Therefore, gage data which are to be carefully compared to remotely sensed rainfall should be spatially interpolated to represent an areal average over the area sampled by the remote sensor. The importance of performing this interpolation will depend on the rainfall gradients and the spatial geometry of the gage network with respect to the underlying landscape. In mountainous terrain, greater confidence can be placed in the gage-based areal average if the interpolation procedure attempts to produce an unbiased estimate by taking expected spatial variability into account. There are a variety of techniques for interpolating point measurements of nonstationary fields; application of many of

these techniques is facilitated by the availability of a description of the nonstationarity. Development of the MSWR model was motivated by a desire to produce such a description for the Southwest.

Below, gage-measured rainfall (interpolated to produce areal values over one-degree cells) is compared to satellite-measured rainfall (averaged over the same one-degree cells). The purpose of this exercise was to demonstrate a technique that could be used to provide regional ground validation data for rainfall measured by the TRMM satellite. Once the WSR-88D doppler radar is fully implemented within the Southwest, it will be possible to produce even higher-quality ground validation data by blending areal gage data with radar data.

1) METHODS

The microwave-based satellite data (see section 1.d.2 for details) have monthly, $0.15^\circ \times 0.15^\circ$ resolution, and were averaged to $1^\circ \times 1^\circ$ resolution prior to comparison with the gage data. Satellite data were also screened for spurious rainfall signals resulting from enhanced surface emissivity over open surface water. Many (but not all) of the cells containing large lakes and reservoirs exhibited unusually high rates of morning rainfall. The amount of spurious morning rainfall was small, however, and did not have a discernible effect on total monthly rainfall.

Rain gage data for specific months were spatially interpolated to give spatial averages over $1^\circ \times 1^\circ$ cells. The interpolation technique was block kriging on detrended data (see Appendix for details). Kriging, which is similar to optimal interpolation, is a linear, minimum variance estimator (Isaaks and Srivastava, 1989; Tabios and Salas, 1985). The kriging estimate is a weighted average of observed values; weights are a function of the spatial geometry of station locations and the spatial covariance structure of the rainfall field. Thus, the method accounts for clustering of stations and the statistical distance between each station and the point/area at which rainfall is to be estimated. Kriging also provides an estimation variance which is based on the rainfall covariance and the location of the stations. Block kriging refers to estimation for an area

rather than a point (using numerical integration). Working with detrended data (deviations from the long term mean) is a way to account for biases which may be introduced by sparse sampling of a field that contains nonrandom spatial trends. The MSWR model was used to evaluate the expected rainfall at each observation station and the expected areal rainfall for each estimation cell. Deviations for each station (observed value minus long-term mean) were interpolated using ordinary block kriging to obtain an area-average deviation, which was then added to the area-average long-term mean for the estimation cell. Kriging was performed using a moving local estimation neighborhood, and a covariance function which was estimated using data from the calibration stations (see Appendix).

The satellite-gage comparison was performed for eight summer months (8/87, 9/87, 7/88, 8/88, 9/88, 7/89, 8/89, and 9/89). These months were selected on the basis of sufficient satellite overpasses (typically 32-41 per month, depending on location) and sufficient rain gage data in our database. The satellite-gage comparison was performed only for those cells that had sufficient gages in or near the cell.

2) RESULTS

The gage-satellite comparison was conducted for monthly values (eight months per one-degree cell) and for mean monthly values (for each cell the average of the eight months).

For monthly values, the correlation between gage-based rainfall and satellite-based rainfall was quite low, although there was little bias (Figure 7 and Table 6). These results are similar to the results of the satellite-gage comparison performed by Negri et al. (1993) for July 1990 in northwestern Mexico (low bias, low correlation, RMSD on the order of the mean). Negri et al. (1993) also showed that monthly rainfall estimates based on a combination of microwave (high-accuracy, low-temporal resolution) and infrared (low-accuracy, high-temporal resolution) data were more accurate than estimates based on microwave data alone.

The comparison between satellite and gage estimates improved when values were averaged over an 8-month period (Figure 7 and Table 6). This suggests that the satellite estimates suffer from random errors (most likely from twice-daily sampling) rather than systematic errors. Lack of consistent bias in the satellite estimates suggests that the gage-satellite comparison should improve with spatial averaging.

Both gage and satellite estimates suffer from sampling problems: insufficient spatial sampling for the gage-based estimate and insufficient temporal sampling for the satellite-based estimate. The kriging estimation variance provides a measure of the sampling uncertainty in the gage-based estimate. Uncertainty in the gage-based estimate was expressed as a confidence region defined by the mean of the kriging estimate $\pm$ twice the kriging standard deviation. In the lower panel of Table 6, statistics were computed considering the distance between the satellite estimate and the *edges* of the gage confidence region. Uncertainty in the gage estimate is reflected by the difference between statistics in the upper panel of Table 6 and statistics in the lower panel; uncertainty in the gage estimate is small in relation to the gage-satellite difference.

2) DISCUSSION

Insofar as satellite rainfall estimates are generally not highly accurate, it is our view that satellite data should be most useful in regions with limited surface data (such as oceans or northern Mexico), for applications concerned with longer time scales, or when blended with surface data. From the viewpoint of regional hydrologic modeling at monthly time scales, the accuracy of the satellite data examined in this study is inadequate. For such modeling, the best possible rainfall estimates would be based, in our opinion, on an optimal combination of gage, ground radar, and satellite data. At the daily time scale, rainfall estimation becomes even more uncertain as the rainfall field becomes even more random and temporal sampling becomes quite poor for some satellite platforms.

4. Summary and Discussion

Within the study area (Arizona, New Mexico, and portions of southern Nevada and California), summer rainfall within a local area can be expected to increase linearly with elevation at a rate of 0.009-0.043 mm of rain per meter of elevation per month. This rate of increase is within the range of rates that have been established for other locations in this study area.

A simple model of long-term mean monthly rainfall (the MSWR model) explains a large part of the spatial variance of observed rainfall. Our experience with the MSWR model was satisfactory, and we recommend that similar models be considered for other rainfall climatology studies. Models of this form (rain = normalized rainfall which varies with location + elevation coefficient * elevation) may be useful for regions with large-scale precipitation trends which cannot be explained by simply regressing precipitation against elevation or latitude/longitude. However, it is possible that models of the MSWR type are best suited for regions where orographic enhancement is primarily due to enhanced convection and that more complicated models are needed where it is important to take into consideration the windward and leeward sides of mountain barriers.

The MSWR model extends prior work on Southwest precipitation-elevation relationships to a much broader area. Furthermore, it offers a refinement to previous knowledge through (a) use of long and consistent periods of record, (b) examination of monthly--as opposed to seasonal--relationships, and (c) estimation of a gridded rainfall climatology which takes orographic factors into account.

Regionally, the rain gage stations used in this study exhibited a modest low-elevation bias. Accounting for the difference between station elevation and mean areal elevation increased the regional estimate of summer rainfall by 9.3%. For individual 2° x 2° cells, however, the difference between rainfall from the MSWR model and the arithmetic mean of gage rainfall was quite significant (ranging from -250% to +41%).

Detrended block kriging (using the MSWR model as a description of the long-term trend) and monthly rain gage data were used to produce unbiased rainfall estimates representing an area-average over $1^{\circ} \times 1^{\circ}$ cells. These gage-based estimates were compared to $1^{\circ} \times 1^{\circ}$ satellite-based rainfall estimates. On a month-by-month basis, there were large differences between the two estimates, although the comparison improved after temporal averaging.

The MSWR model of mean rainfall provides insights into the summer rainfall climatology. When rainfall is normalized to sea level, the northern tip of the Mexican monsoon maximum can be seen in southcentral Arizona. Furthermore, a rainfall "trough" running north-northwest in western New Mexico provides information about dominant moisture sources, i.e., rainfall to the west of this trough is most strongly associated with Mexican monsoon moisture from the Gulf of California/tropical eastern Pacific, while rainfall to the east of this trough is most strongly associated with the Gulf of Mexico.

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5. Appendix: Interpolation Method

For the detrended block kriging technique, areal rainfall over area A was estimated using:

$$\bar{P} = \bar{P}^m + \sum_{i=1}^n w_i P_i^d$$

$$P_i^d = P_i^o - P_i^m \quad (3)$$

where:

$\bar{P}$ = areal-average rainfall estimated for a particular month and year;

$\bar{P}^m$ = areal long-term mean rainfall over area A from MSWR model;

n = number of gages in the local neighborhood (estimation cell and adjacent cells);

w_i = kriging weight for gage i ;

P_i^d = detrended rainfall at gage i for a particular month and year;

P_i^o = observed rainfall at gage i for a particular month and year; and

P_i^m = long-term mean rainfall at gage i from MSWR model.

The kriging weights w_i were obtained by solving the ordinary block kriging equations:

$$\sum_{i=1}^n C_{ij} w_i + \lambda = \bar{C}_{jA} \quad \text{for } j = 1 \text{ to } n$$

$$\sum_{i=1}^n w_i = 1 \quad (4)$$

where λ is a Lagrange parameter, and C_{ij} is the covariance between rainfall at station i and rainfall at station j . $\bar{C}_{jA}$ is the point to block covariance:

$$\bar{C}_{jA} = \frac{1}{|A|} \sum_{i|i \in A} C_{ij} \quad (5)$$

The kriging estimate is composed of the mean $\bar{P}$ (given by Equation 3) and a variance:

$$\sigma^2 = \bar{C}_{AA} - \lambda - \sum_{i=1}^n w_i \bar{C}_{iA} \quad (6)$$

where $\bar{C}_{AA}$ is the mean covariance between pairs of locations within A.

The covariance structure of monthly rainfall was estimated using data from the calibration stations. Exponential, spherical, and Gaussian covariance models were fitted. The exponential model was selected based on a least squares criterion (Figure 8). The correlation scale (distance beyond which there is no autocorrelation) was 1120, 1170, and 1285 km for July, August, and September, respectively. Examination of maps of rainfall anomalies (deviation from long-term mean) for specific months suggests that anomalies tend to cover part of the study area rather than the entire Southwest.

For the purpose of computing the kriging variance (Equation 6), the covariance function was scaled by the variance of the observations within the local search neighborhood. This allows the variance of a particular estimate to be related to the variability of the relevant observations.

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Table 1. Normalized Rainfall Parameter *B* of the MSWR Model. Values are in mm; there is one value per 1° x 1° cell.

(a) Normalized Rainfall for July

	116- 117° W	115- 116°W	114- 115°W	113- 114°W	112- 113°W	111- 112°W	110- 111°W	109- 110°W	108- 109°W	107- 108°W	106- 107°W	105- 106°W	104- 105°W
36-37°N	3	5	5	-31	-41	-48	-46	-44	-51	-44	-36	-46	-23
35-36°N	2	5	1	-26	-23	-29	-36	-41	-49	-32	-23	-19	2
34-35°N	1	-3	-3	8	1	-1	-12	-22	-29	-22	-26	0	6
33-34°N	2	2	1	6	7	8	10	-9	-7	-21	-16	-1	-4
32-33°N	1	2	2	7	10	14	13	3	-2	-6	-1	11	-1
31-32°N	0	1	1*	11	11	61	40	29	8	-12	-14	-14	-1

(b) Normalized Rainfall for August

	116- 117° W	115- 116°W	114- 115°W	113- 114°W	112- 113°W	111- 112°W	110- 111°W	109- 110°W	108- 109°W	107- 108°W	106- 107°W	105- 106°W	104- 105°W
36-37°N	3	5	5	-21	-29	-40	-39	-37	-44	-38	-31	-45	-22
35-36°N	2	5	5	-13	-12	-21	-30	-35	-34	-25	-24	-19	4
34-35°N	2	5	5	27	17	11	-1	-17	-19	-22	-27	7	11
33-34°N	4	9	13	20	18	19	23	-5	-2	-17	-15	-6	2
32-33°N	2	8	11	22	25	21	16	5	3	-1	-6	3	5
31-32°N	1	5	5*	33	33	54	38	24	8	-10	-19	-19	5

(c) Normalized Rainfall for September

	116- 117°W	115- 116°W	114- 115°W	113- 114°W	112- 113°W	111- 112°W	110- 111°W	109- 110°W	108- 109°W	107- 108°W	106- 107°W	105- 106°W	104- 105°W
36-37°N	3	3	3	-9	-8	-10	-17	-15	-10	-8	-5	-15	-3
35-36°N	1	3	4	-7	-1	-3	-8	-9	-13	-5	-3	2	9
34-35°N	4	5	5	17	11	13	7	-2	-4	2	-3	9	15
33-33°N	7	8	9	13	14	16	15	8	10	2	6	12	18
32-33°N	5	9	11	11	11	15	12	8	10	9	10	12	31
31-32°N	4	7	7*	11	11	19	14	9	7	7	7	7	31

* Uncertain estimate due to insufficient local data

Table 2. Elevation Coefficients of the MSWR Model.

Month(s)	Elevation Coefficient ^a
<i>Main Portion of the Study Area (Right of Dashed Line in Figure 2)</i>	
July and August	0.043254
September	0.020757
<i>Dry Region (Left of Dashed Line in Figure 2)</i>	
July, August, and September	0.0085677

^a mm of rain per m of elevation per month

Table 3. Comparison of Point Rainfall Observations and Point Predictions of the MSWR Model.

	n	bias (mm)	RMSD (mm)	r ²
<i>Calibration</i>				
July	91	-0.1	8.1	0.92
August	91	-0.4	8.4	0.91
September	91	-0.2	4.5	0.90
July, Aug., and Sept.	273	-0.2	7.3	0.92
<i>Validation</i>				
July	26	0.1	12.9	0.64
August	26	-0.3	12.7	0.62
September	26	0.3	6.9	0.60
July, Aug., and Sept.	78	0.1	11.2	0.70

RMSD = Root Mean Square Difference

Table 4. Gridded Long-Term Mean Rainfall. Values are areal averages in mm for 1° x 1° cells.

(a) July Rainfall

	116- 117°W	115- 116°W	114- 115°W	113- 114°W	112- 113°W	111- 112°W	110- 111°W	109- 110°W	108- 109°W	107- 108°W	106- 107°W	105- 106°W	104- 105°W
36-37°N	12	16	12	9	37	25	36	36	26	48	70	65	63
35-36°N	8	14	9	15	57	44	42	48	42	58	67	76	70
34-35°N	8	3	22	57	64	77	65	63	68	60	56	78	67
33-34°N	21	18	16	29	32	52	77	84	91	59	56	76	57
32-33°N	7	3	10	19	34	46	68	61	71	58	60	85	48
31-32°N	5	13	4*	19	38	108	102	93	69	42	42	45	49

(b) August Rainfall

	116- 117°W	115- 116°W	114- 115°W	113- 114°W	112- 113°W	111- 112°W	110- 111°W	109- 110°W	108- 109°W	107- 108°W	106- 107°W	105- 106°W	104- 105°W
36-37°N	12	16	12	19	49	33	43	43	33	54	75	66	64
35-36°N	8	14	13	28	68	52	48	54	57	65	66	76	72
34-35°N	9	11	30	76	80	89	76	68	78	60	55	85	72
33-34°N	23	25	28	43	43	63	90	88	96	63	57	71	55
32-33°N	8	9	19	34	49	53	71	63	76	63	55	77	54
31-32°N	6	17	11*	41	60	101	100	88	69	44	37	40	55

(c) September Rainfall

	116- 117°W	115- 116°W	114- 115°W	113- 114°W	112- 113°W	111- 112°W	110- 111°W	109- 110°W	108- 109°W	107- 108°W	106- 107°W	105- 106°W	104- 105°W
36-37°N	12	14	10	13	30	25	22	24	27	36	46	38	38
35-36°N	7	12	12	16	38	32	29	34	31	38	40	47	42
34-35°N	11	11	17	41	41	51	44	39	43	41	36	46	44
33-34°N	18	15	16	24	26	37	47	52	57	40	40	49	43
32-33°N	11	10	15	17	23	3	39	36	45	40	39	47	54
31-32°N	9	14	12*	15	24	41	44	39	36	33	34	35	55

* Uncertain estimate due to insufficient local data

Table 5. Bias in Arithmetic Estimates of Mean Summer Rainfall ($2^{\circ} \times 2^{\circ}$ cells). Values are the percent difference between MSWR rainfall and the arithmetic average of gage rainfall; negative values indicate that the arithmetic mean is an underestimate.

	116- 118°W	114- 116°W	112- 114°W	110- 112°W	108- 110°W	106- 108°W	104- 106°W
35-37°N	-250%	-20%	+41%	+19%	-14%	-28%	-11%
33-35°N	-195%	-128%	0%	-36%	-14%	-49%	-6%
31-33°N	+13%	-162%	-32%	-13%	-2%	-24%	+25%

Table 6. Satellite-Based Rainfall Minus Kriging-Estimated Rainfall (1° x 1° areas).

MONTHLY VALUES <i>(eight months per cell)</i>			MEAN MONTHLY VALUES <i>(mean over the eight months)</i>		
Bias	RMSD	r^2	Bias	RMSD	r^2
<i>Comparison against the mean of the gage estimate</i>					
-7.9 mm	52.9 mm	0.10	-7.8 mm	21.1 mm	0.40
<i>Comparison which considers uncertainty in the gage estimate</i>					
-5.2 mm	45.3 mm	-	-	-	-

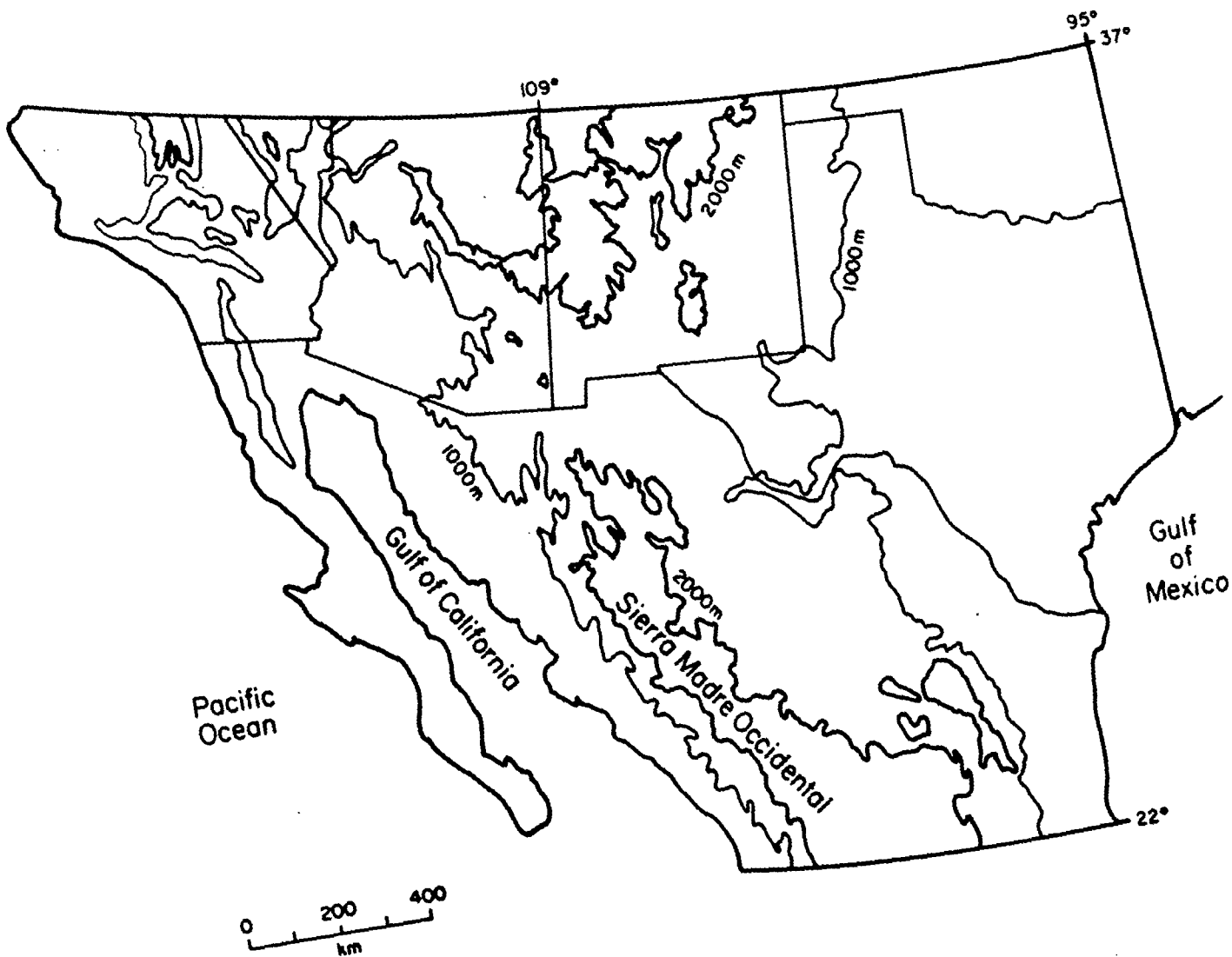


Fig 1

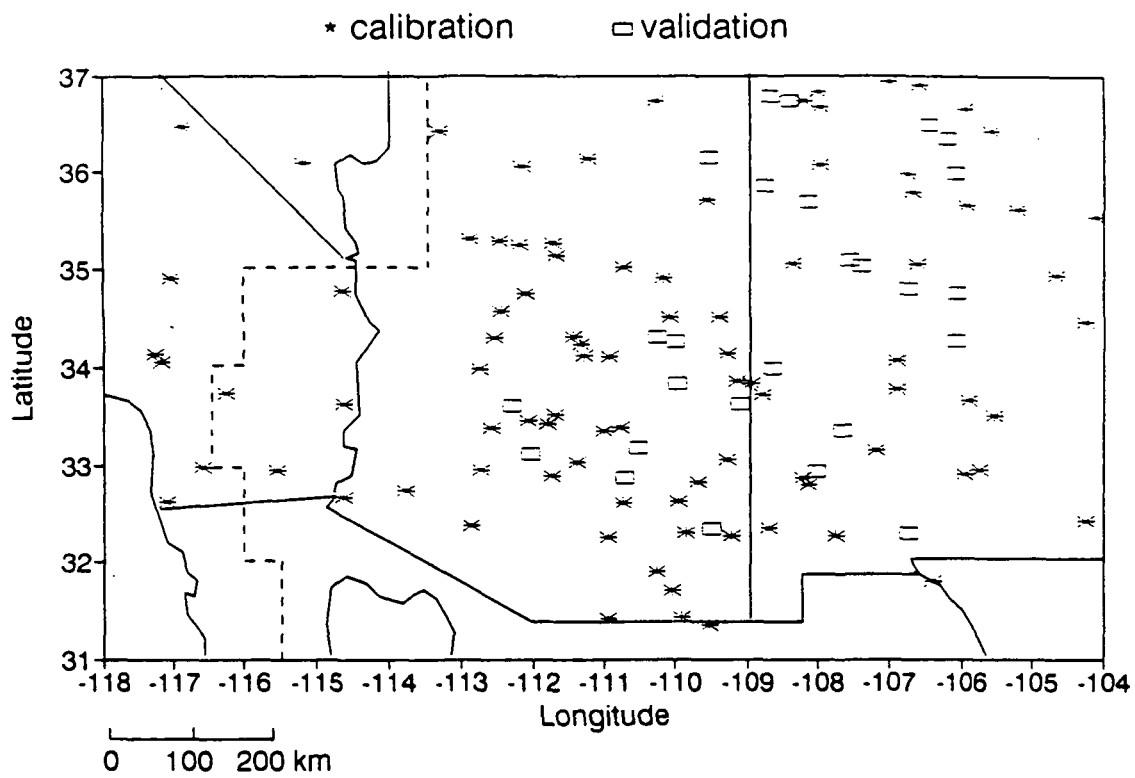


Fig 2

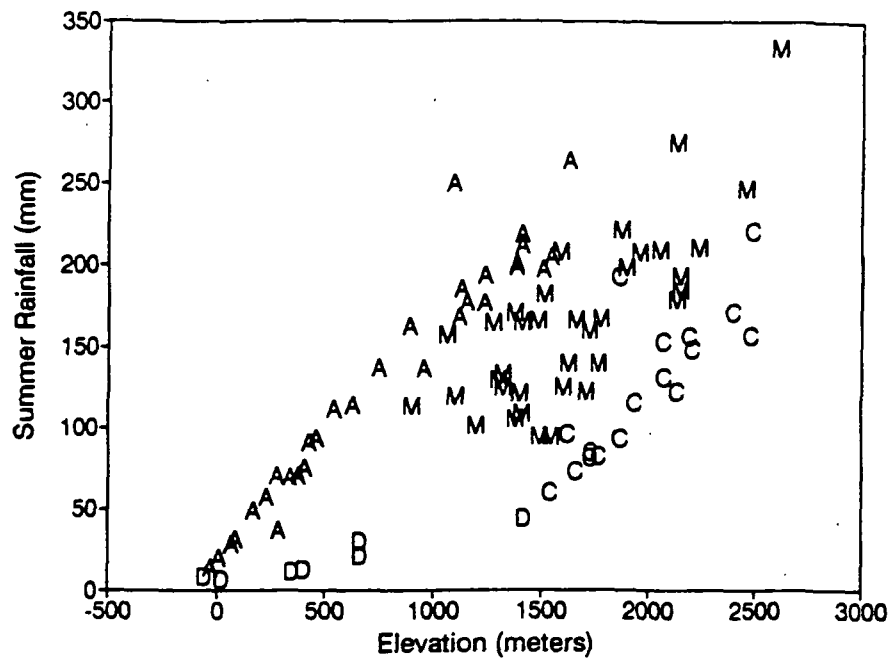


Fig. 3

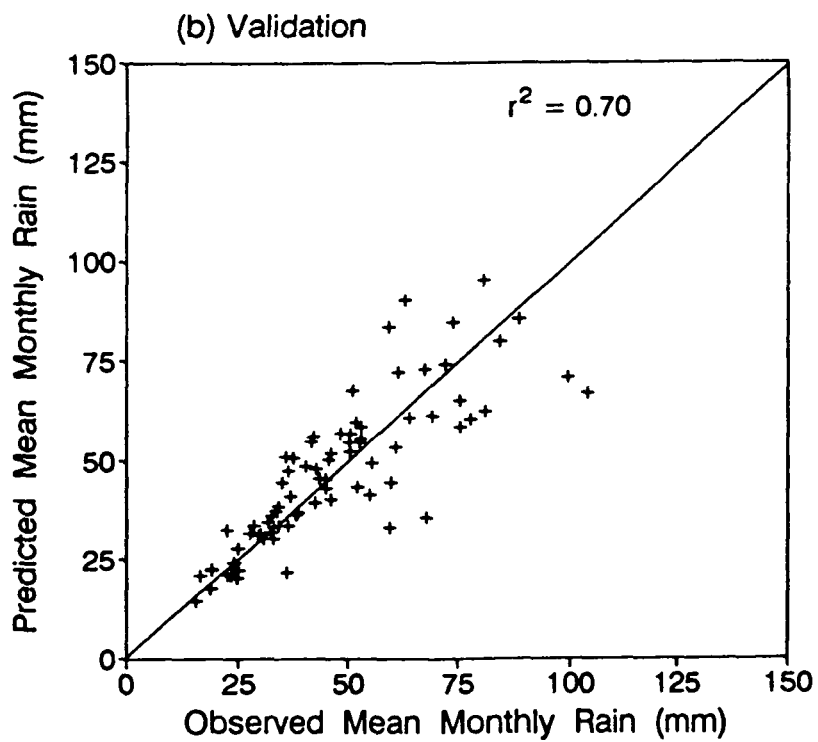
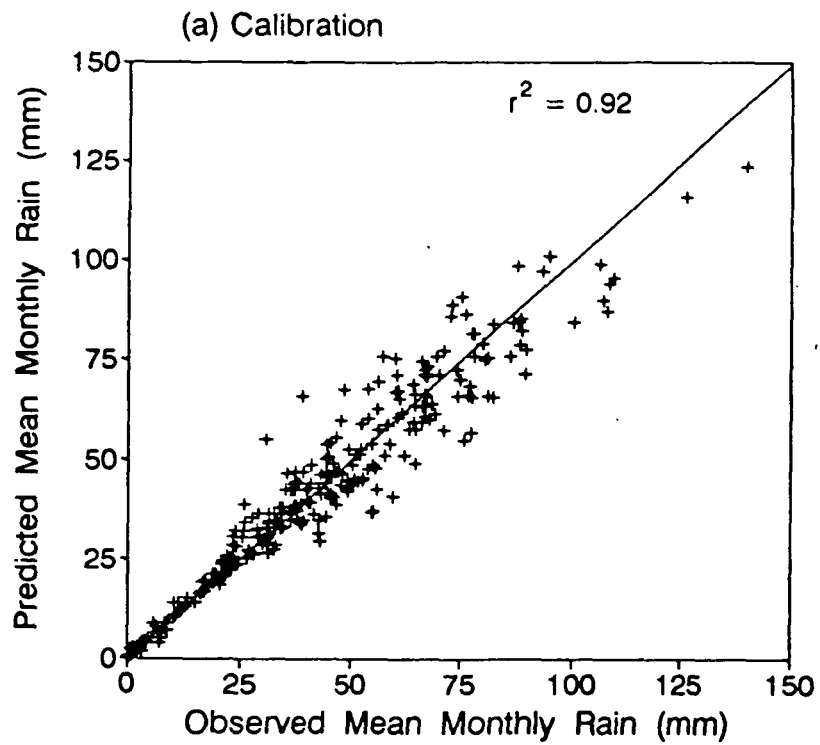


Fig 4

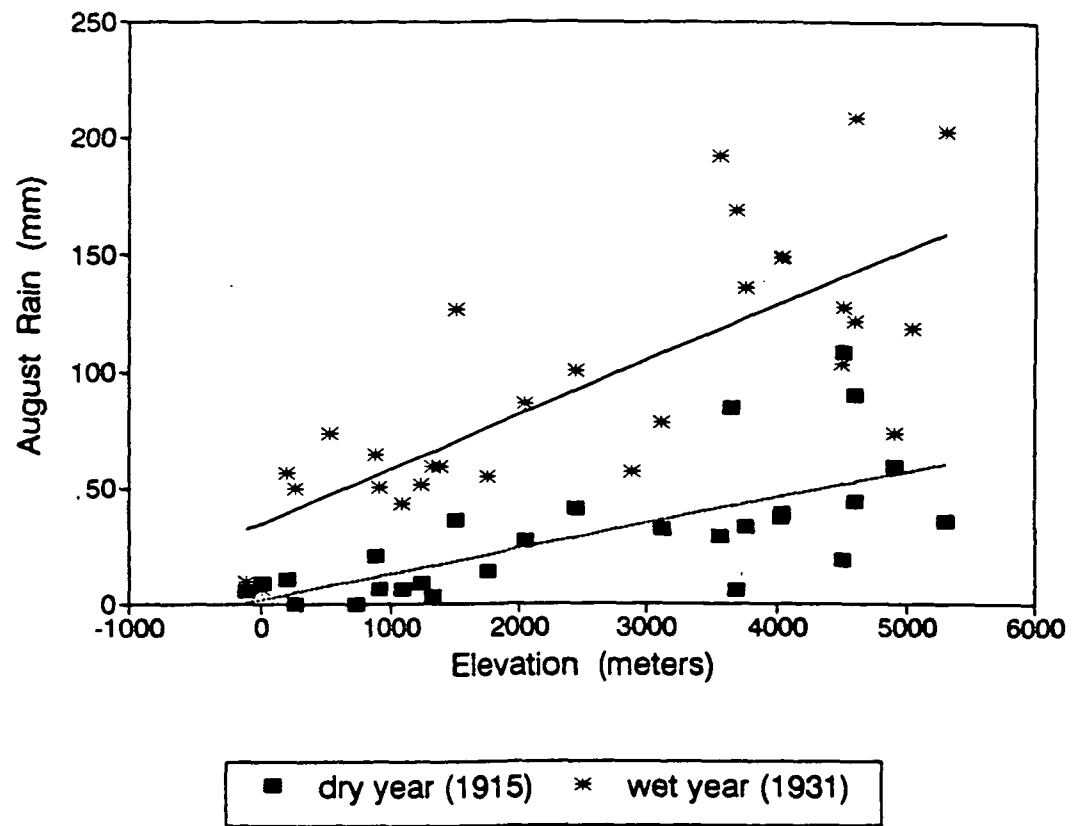
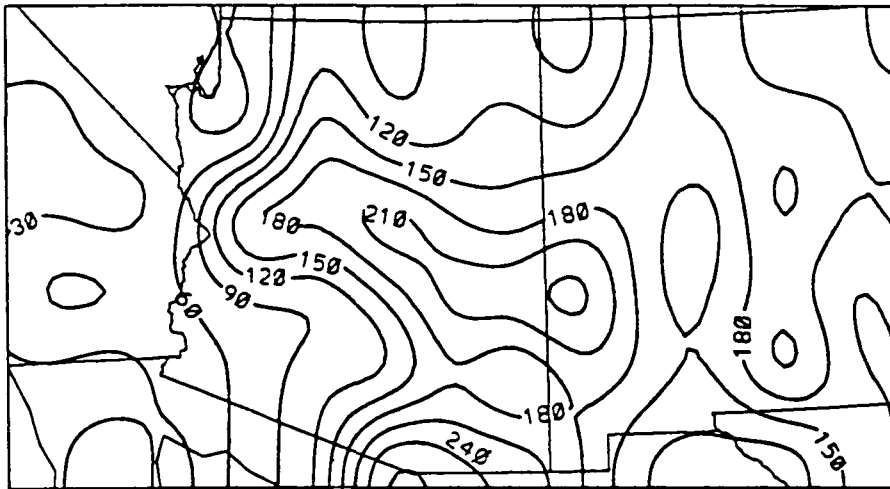


Fig. 5

(a) with elevation



(b) normalized to sea level

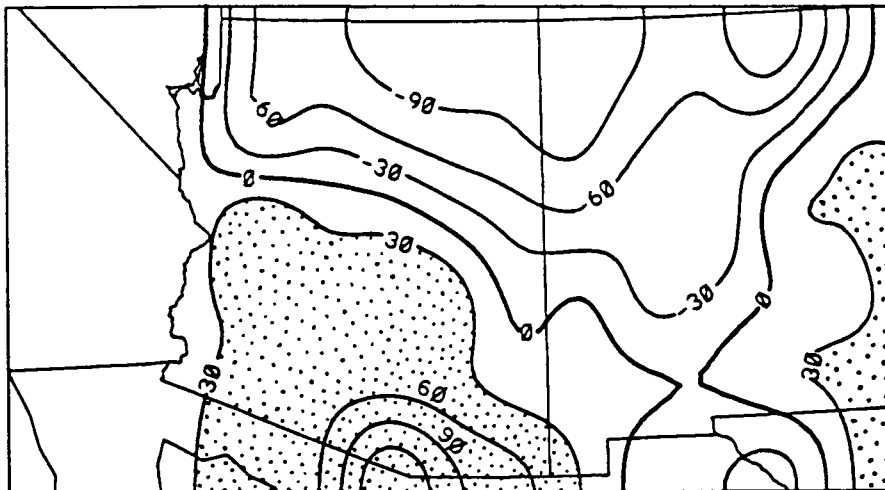


Fig.6

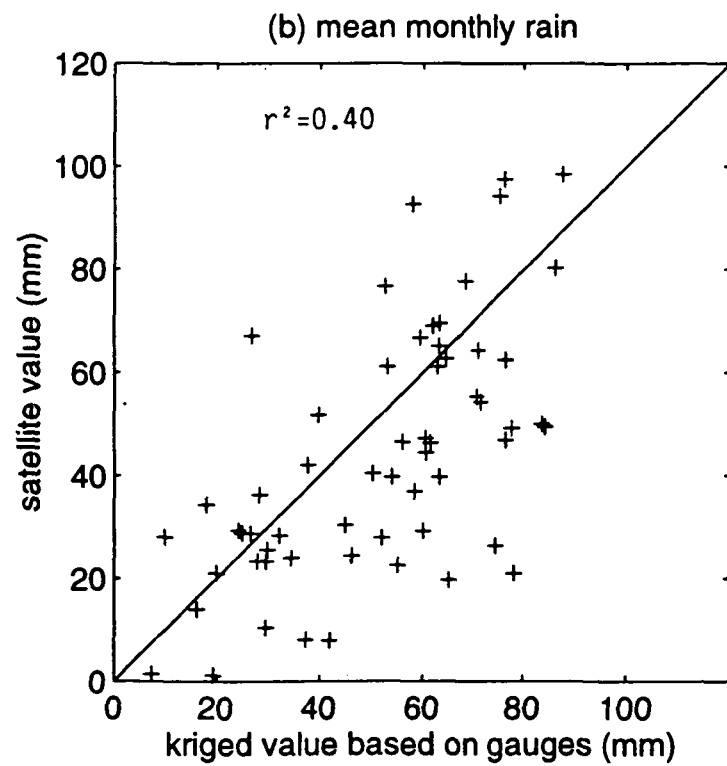
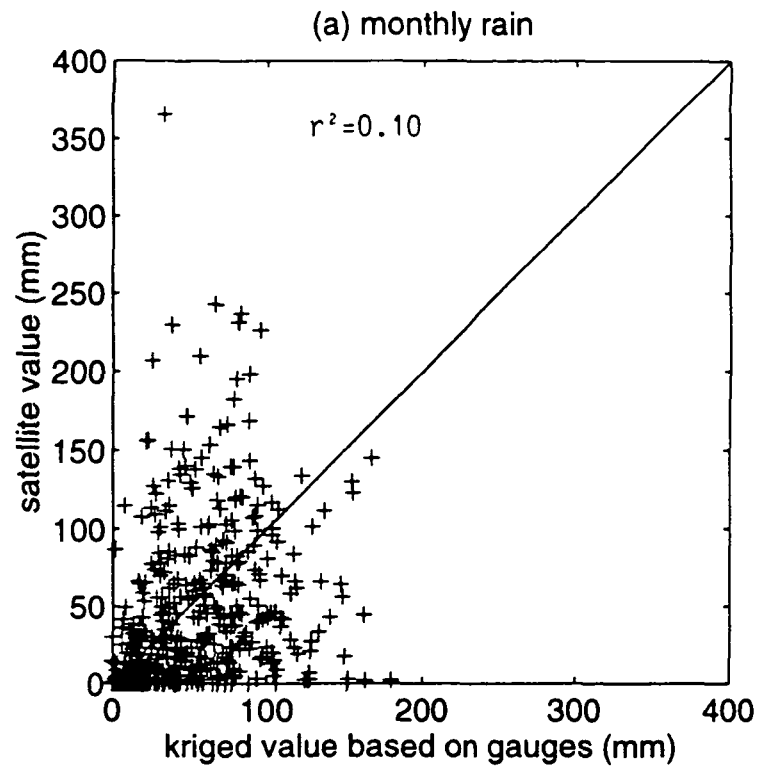


Fig 7

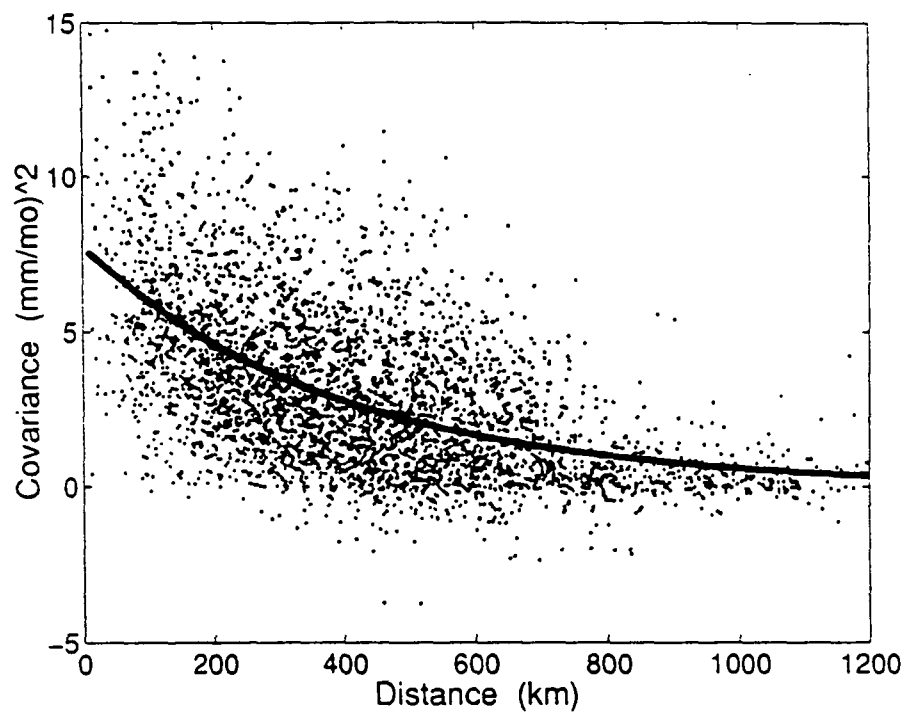


Fig. 8

APPENDIX IV — Water Vapor Transport Associated with the Summertime North American
Monsoon as Depicted by ECMWF Analyses

**Water Vapor Transport Associated with the Summertime
North American Monsoon as Depicted by ECMWF Analyses**

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Abstract

The origins and transport of water vapor into the semi-arid Sonoran Desert region of southwestern North America are examined for the July-August wet season. Vertically-integrated fluxes and flux divergences of water vapor are computed for the 8 summers 1985-1992 from ECMWF mandatory-level analyses possessing a spectral resolution of triangular 106 (T106).

The ECMWF analyses indicate that transports of water vapor by the time-mean flow dominate the transports by the transient eddies. Most of the moisture at upper-levels (above 700 mb) over the Sonoran Desert arrives from over the Gulf of Mexico, while most moisture at low-levels (below 700 mb) comes from the northern Gulf of California. There is no indication of moisture entering the Sonoran Desert at low-levels directly from the southern Gulf of California or the tropical East Pacific. Water vapor from the tropical East Pacific can enter the region at upper-levels after upward transport from low-levels along the western slopes of the Sierra Madre Occidental mountains of Mexico and subsequent horizontal transport aloft.

The T106 ECMWF analyses, when only the mandatory-level analyses are used, do not possess sufficient precision to yield accurate estimates of highly differentiated quantities such the divergence of the vertically-integrated flux of water vapor. Even at a T106 resolution, the northern Gulf of California and the terrain of the Baja California peninsula are not adequately resolved.

Introduction

As the boreal solstice approaches, the North Pacific High builds off the West Coast of North America, relinquishing much of its influence on the Sonoran Desert region of the SW United States and NW Mexico. (Figure 1 shows a map with the geographical regions referred to in this paper.) Simultaneously, the Bermuda High thrusts northwestward into the region, and the time-mean mid-tropospheric (500-700 mb) flow backs from WNW during May and June to SSE during July and August. Concurrent with the wind shift, precipitable water and convective activity over the region increase dramatically. Consideration of the change in the prevailing wind and the geographical distribution of land-sea boundaries led many earlier researchers (Reed 1933; Jurwitz 1953; Bryson and Lowry 1955; Reitan 1957) to conclude that the summertime moisture over the Sonoran Desert is transported from the Gulf of Mexico, along the western flank of the Bermuda High.

An analysis of the water vapor transport over North America by Benton and Estoque (1954), however, suggested another source of summertime moisture situated to the west of the continental divide, separate from the larger flux from the Gulf of Mexico. More evidence of a Pacific source of moisture came from Rasmusson (1967) whose analysis showed that water vapor east of the divide clearly originates from the Gulf of Mexico/Caribbean Sea while moisture over the Sonoran Desert appears to originate from the Gulf of California. The issue of moisture from the tropical East Pacific was not resolved because neither analysis extended into Mexico.

Hales (1972, 1974) proposed that moisture over the Sonoran Desert comes in the

form of short-lived, low-level surges up the Gulf of California. Brenner (1974) concurred, adding that these surges appear to be independent from the large-scale circulation. Both Hales and Brenner expressed skepticism that moisture from the Gulf of Mexico could pass over the Sierra Madre Occidental range and still make significant contributions to the precipitable water over the Sonoran Desert. Sellers and Hill (1974), on the other hand, maintained that monsoon moisture comes primarily from the Gulf of Mexico.

Reyes and Cadet (1986, 1988) examined moisture flux over the tropical Americas during the period May-August 1979. They proposed that the intensification of the South Pacific anticyclonic gyre propels low-level moisture across the equator towards western Mexico. Once there, it could reach the Sonoran Desert either at low-levels as a result of southeasterly gulf-surges or at mid-levels after convective mixing with midtropospheric moisture from the east and subsequent transport northward around the western flank of the subtropical ridge. Their description stressed the importance of both Gulf of Mexico and tropical East Pacific moisture sources and the coupling of the two through convective transport.

Recent research using data from special field programs provides further evidence for a low-level flux of water vapor along the Gulf of California (Badan-Dangon *et al.* 1991; Douglas *et al.* 1993). Contrary to the short-lived surges proposed by Hales (1972, 1974) and Brenner (1974), these studies reveal a persistent transport of moisture along the Gulf of California by the time-mean wind. Douglas *et al.* (1993) also show that relatively dry air at mid-levels is advected from east of the Sierra Madre Occidental toward western Mexico, and they argue that most of the moisture over the Sonoran

Desert comes from the tropical East Pacific Ocean or directly off the Gulf of California.

To this day, two fundamental issues remain unresolved: (1) the relative importance of the Gulf of Mexico, the Gulf of California, and the eastern Tropical Pacific as moisture sources for the Sonoran Desert, and (2) the primary means by which the moisture is transported into the region. The issues are of more than academic interest, they also have operational implications if seasonal precipitation forecasts for the region during the summer are ever to be realized. Hence, we believe that the need exists to examine these issues further.

To provide insight into these questions, we desire to diagnose the moisture flux and water vapor balance over SW North America and adjacent ocean environs for the July-August period using a multi-year, high resolution data set produced by a modern-day atmospheric forecast/analysis system. Although the Desert Southwest has been included in many prior investigations of the transport and balance of water vapor, most were based on only a single month, season, or year of data (Benton and Estoque 1954; Starr and Peixoto 1958; Starr *et al.* 1965; Reyes and Cadet 1988; Trenberth 1991). The question of representativeness arises in view of the interannual variability that characterizes the region (Carleton *et al.* 1990). A few multi-year climatologies that included SW North America exist, but these often concerned themselves with differences among the annual means (e.g. Rosen *et al.* 1978). Even when winter and summer regimes were explicitly contrasted, conditions for June, July, and August are usually averaged to represent the summer mean (Oort 1983; Peixoto and Oort 1992; Roads *et al.* 1994). Because June conditions are so drastically different from those of July and

August (Douglas *et al.* 1993), its inclusion would distort the flow patterns of “wet” phase of the monsoon. Moreover, all of these studies relied either on radiosonde data which effectively ignores the adjacent oceanic regions, or on gridded data sets having resolutions much too coarse to resolve the Gulf of California, a postulated source of moisture for the Sonoran Desert.

As a means to address the problem using modern-day atmospheric analyses possessing finer spatial and higher temporal resolutions than in previous studies, this work utilizes the global analyses from the European Center for Medium-range Weather Forecasting (ECMWF) to compute an 8 year climatology of the water vapor flux and balance associated with the summer phase of the North American Monsoon. To our knowledge, no such analysis *focusing on the semi-arid Sonoran Desert* has been done using ECMWF analyses.

This paper is organized as follows. Sections 2 and 3 describe the ECMWF data set and the methodology, respectively. Section 4 gives a brief overview of the climatology for the region of interest. Results for the transport of water vapor and its flux divergence are presented in Section 5. Conclusions and future research directions are discussed in Section 6.

Data

Uninitialized global analyses produced by the ECMWF are used in this study. The analyses were obtained from the National Center for Atmospheric Research (NCAR). Analyses are available four times daily (0000, 0600, 1200, and 1800 UTC) for

the surface and 14 mandatory levels from 1000 mb to 10 mb (the recently added 925 mb level is not included). The spectral truncation for the ECMWF analyses is triangular 106 (T106).

The spectral coefficients for all available analyses for the period July and August, 1985 to 1992, are first transformed to a Gaussian grid having a spacing of approximately 1.125° latitude by 1.125° longitude. In order to limit the data storage requirements and allow for processing on local workstations, only a subset of the globe covering the region of interest, bounded approximately (to the nearest five degrees) by 5°N to 50°N and 75°W to 130°W , inclusive, was transferred to the University of Arizona for analysis. Since virtually all of the water vapor in the atmosphere occurs below 200 mb, only analyses for the surface and eight lowest mandatory levels (1000, 850, 700, 500, 400, 300, 250, 200 mb) are included in the calculation.

Because the ECMWF analyses are operational, they undergo continuous changes in the data analysis procedures. These changes can produce temporal trends and discontinuities which could impact long-term climatic means and eddy statistics (e.g. Trenberth and Olson, 1988). Moreover, the divergent wind component, vertical velocity, and moisture, the three crucial quantities required to compute an accurate moisture budget, are the fields most adversely affected (Trenberth and Olson 1988). Furthermore, over data sparse regions, such as the oceans and to a lesser degree over Mexico itself, analysis quantities are largely determined by the first-guess forecast fields even at the synoptic hours of 0000 UTC and 1200 UTC. Clearly the validity of any conclusions drawn using the ECMWF analyses depends critically on their integrity.

While it is very difficult, if not impossible, to determine precisely the fidelity of the ECMWF analyses, estimates of the uncertainty can be obtained. The uncertainty in the derived water vapor budgets and moisture fluxes can be estimated assuming the wind and specific humidity at each grid point have an accuracy comparable to radiosonde data. This clearly denotes a lower bound estimate of the uncertainty. Thus, whenever a budget value is less than that due to the analysis uncertainty, the budget value must be considered insignificant. A crude consistency check can also be made by comparing our results for the T106 ECMWF analyses to previous results based solely on the analysis of radiosonde data.

Analysis Procedures

Vertically-Integrated, Atmospheric Flux of Water Vapor

To illustrate the water vapor transport for the monsoonal period, vertically-integrated moisture flux vectors ($\vec{Q}$) can be computed:

$$\vec{Q} = \int_{p_{\text{top}}}^{p_{\text{sfc}}} (q \vec{V}) \frac{dp}{g} \quad (1)$$

where q is specific humidity, $\vec{V}$ is the horizontal wind vector, p_{sfc} is the surface pressure and p_{top} is the pressure at the top of the atmosphere. It is also of interest to compare the transport at low levels to that at higher levels. To examine this, the moisture flux for the atmospheric layers above and below 700 mb, and the exchange of water mass between them, are considered. In this case, transport vectors are computed by integrating (1) from the surface to 700 mb, and from 700 mb to 200 mb, respectively.

The exchange of moisture between these two layers is represented by the vertical flux through the 700 mb level.

Atmospheric Water Vapor Balance

The general balance equation for atmospheric water vapor can be expressed as

$$E - P = \frac{\partial W}{\partial t} + \nabla \cdot \bar{Q} - \frac{\omega_{\text{top}} q_{\text{top}}}{g}, \quad (2)$$

where ω is the pressure vertical velocity, E is evaporation per unit area, P is precipitation per unit area, and W is the precipitable water defined by

$$W = \int_{p_{\text{top}}}^{p_{\text{sf}}} q \frac{dp}{g}. \quad (3)$$

All other symbols have their usual meteorological meaning. A complete derivation of the general water vapor balance equation is given in Appendix 1. Equation (2) states that the difference between evaporation and precipitation at the earth's surface equals the sum of the local change in precipitable water, the divergence of the vertically integrated horizontal water vapor flux and the vertical water vapor transport through the top of the atmosphere. The last term of eq. (2) is negligible since $q \approx 0$ at 200 mb.

Averaging (2) over time yields

$$\nabla \cdot \bar{Q} \approx \bar{E} - \bar{P}, \quad (4)$$

where an overbar denotes a time average over the period July-August. The storage term

can be neglected for an averaging period of two months since it is typically much smaller than the mean flux divergence (Starr and Peixoto 1958; Rasmusson 1966; Palmén 1967; Peixoto 1973). Equation (4) also neglects horizontal diffusion and the horizontal transport of liquid and solid phases. Although these are important elements of the water balance for individual convective elements, they are both generally much smaller than the remaining terms when considering long averaging periods such as a month or more, and/or large spatial extents (Starr and Peixoto 1958; Palmén 1963; Peixoto 1973; Rasmusson 1968, 1977) such as a 1.125° by 1.125° grid.

Equation (4) provides an estimate of the mean water balance for a column of air extending from the surface to the top of the atmosphere, or 200 mb in this case. Areas with an excess of evaporation over precipitation ($E - P > 0$) are termed water vapor source regions, while areas with an excess of precipitation over evaporation ($E - P < 0$) are termed sink regions.

Regional Balance of Water Vapor and Moisture Transport Across Regional Boundaries

Spatially-averaged water vapor budgets can be obtained by integrating (4) over an area A giving

$$\frac{1}{A} \int \int_A \nabla \cdot \bar{\mathbf{Q}} dA = \{\bar{E} - \bar{P}\}$$

which, with the aid of Gauss's Theorem, can be written as

$$\oint_C \bar{\mathbf{Q}} \cdot \hat{n} dC = \{\bar{E} - \bar{P}\} , \quad (5)$$

where $\{ \}$ denotes an areal average, $\hat{n}$ is the outward unit vector normal to the regional boundary, $\overline{\vec{Q}} \cdot \hat{n}$ is the moisture transport across the regional boundary, and the line integral is computed along the closed path delimiting the region.

Miscellaneous Numerical Procedures

Vertical integrals are calculated by the trapezoid rule, with all fields assumed to vary linearly with pressure between mandatory levels. Spatial derivatives are computed using centered finite differences on the 1.125° grid. This procedure is not consistent with the ECMWF formulation which operates on the spectral coefficients to obtain horizontal derivatives. Since only a subset of the hemisphere was obtained, we are unable to use spectral processing. As will become clear later, the numerical error introduced by this procedure is negligible compared to other uncertainties and error sources inherent in our analysis.

Time-mean quantities are obtained by averaging over all 8 summers and all four analysis times (0000, 0600, 1200 and 1800 UTC). Transient eddy statistics are obtained as departures from the 8-year seasonal means at the four daily analysis times, then averaging those results over all four analysis times. Thus interannual variations are contained in the eddy statistics, but contributions from diurnal cycle excluded. The seasonal cycle is not removed from the data.

Overview of the Regional Climatology for July-August

In this section, the mean July-August wind and moisture fields over SW North American as depicted by the T106 ECMWF analyses are briefly described and compared

with results from earlier studies based on other data sources and years. Our primary purpose is not to present a detailed discussion of the regional climate; such information can be found elsewhere (e.g. Douglas and Reyes 1993). Rather, we desire to provide a proper background in which to interpret the moisture fluxes and budgets and to demonstrate the fidelity of ECMWF analyses.

The large-scale, low-level flow over the region is strongly influenced by the Pacific and Atlantic Subtropical Highs (Fig. 2b), with brisk southerlies over Texas and northwesterlies west of Baja California. Evidence of the thermal low can be seen in the cyclonic winds over the lower Colorado River Valley. The winds at 500 mb (Fig. 2a) are characterized by easterlies over the tropics and an anticyclonic circulation centered over southern New Mexico. The upper-level ECMWF winds agree closely with prior analysis of time-mean radiosonde data (e.g. Douglas and Reyes 1993). The surface and low-level ECMWF winds are also consistent except over the northern Gulf of California, and even in that region the ECMWF analyses denote an improvement over other global objective analyses. Analysis of special field observations (Badan-Dangon *et al.* 1991; Douglas and Reyes 1993) reveals that light southerlies to south-southeasterlies mark the time-mean, low-level wind field over the Gulf of California. NMC analyses, for example, place low-level northwesterlies winds over the region (Stensrud *et al.* 1995). The ECMWF analyses, while not totally consistent with the special field observations, do at least yield a southerly component over the region. The reasons for the erroneous wind direction are not known, but as we later discuss it may be a ramification of a horizontal resolution which is still too coarse to adequately resolve the Gulf of California and the surrounding

terrain.

A major advantage with using modern objective analyses is that the vertical motion fields, consistent with the model formulations and data assimilation procedures, are available. The time-mean, vertical velocity field at 500 mb (Fig. 3) indicates localized ascent over the southern Rockies and the Sierra Madre Occidental with weaker, more widespread descent situated over the southern Central Plains States, the East Pacific Ocean, the Gulf of California, and the Gulf of Mexico. The region of upward motion is consistent with precipitation (Douglas *et al.* 1993; Negri *et al.* 1994) and infrared cloud climatologies (Douglas *et al.* 1993), while the overall distribution of vertical velocity is in qualitative agreement with the June, July, and August mean determined by Oort (1983). The ascent over the Sierra Madre Occidental in the ECMWF analyses is much stronger than that given by Oort (1983), however.

The distribution of specific humidity at the surface (Fig. 2b) strongly reflects the underlying terrain, with high values flanking the southern Rockies and the Mexican Plateau. Surface humidity increases eastward into the Central United States, and a moist tongue is evident over the Gulf of California and the coast of western Mexico. In the middle troposphere (Fig. 2a), a band of enhanced moisture extends northward from the tropical East Pacific into southern Mexico before curving anticyclonically into Arizona and New Mexico. The 500 mb wind and moisture fields indicate that the easterly winds over the Gulf of Mexico advect drier air into western Mexico as discussed earlier by Douglas *et al.* (1993). The precipitable water (Fig. 4) indicates that deep moisture exists over western Mexico and the tropical East Pacific, with values being

larger over that region than anywhere else in the domain. Similar moisture distributions have been previously reported (Starr *et al.* 1965; Hales 1974; Hagemeyer 1991; Douglas *et al.* 1993; Negri *et al.* 1994).

In summary, the mean distributions of the ECMWF wind, vertical velocity, and moisture are qualitatively consistent with results from prior studies. With the noted exception of surface winds over the northern Gulf of California, the T106 ECMWF analyses also appear to give quantitatively accurate, time-mean fields over the domain.

Results

Vertically-Integrated, Atmospheric Flux of Water Vapor

The distribution of mean flux vectors of water vapor, integrated from the surface to 200 mb (SFC-200 mb), exhibits several noteworthy features over the SW United States and NW Mexico (Fig. 5). On the whole, the SFC-200 mb flux vectors bare close resemblance to the surface wind field (cf. Fig. 5 and 2b), a result that reflects the much larger specific humidities at low-levels. Also as expected, the SFC-200 mb flux vectors over the Mexican Plateau are noticeably smaller than neighboring regions, which reflects the impact of underlying high terrain on reducing the limits of the vertical integration.

The strongest flux vectors curve anticyclonically from the Gulf of Mexico into the South Central Plains. A weaker southerly transport is also apparent over the northern Gulf of California and western Arizona. These two moisture streams were first described by Rasmusson (1967). A third feature of interest is a southeasterly flux off the southwest coast of Mexico that emanates from the tropical East Pacific. This moisture plume from

the tropical East Pacific, unlike the two described by Rasmusson (1967), appears to flow no farther north than $\sim 25^{\circ}\text{N}$, as the flux vectors veer to the west over the southern Gulf of California. Based solely on the distribution of the SFC-200 mb fluxes, it appears that moisture over the Sonoran Desert comes predominately from the Gulf of Mexico and northern Gulf of California, with little or no input from the tropical East Pacific.

A comparison of the time-mean flow and transient eddy contributions to the moisture transport (Figs. 6a and b, respectively) reveals a dominance by the time-mean circulation. While the net transport by the transients is minor compared to that by the time-mean flow, it would be premature to conclude that transient fluctuations are unimportant to the regional water vapor balance. As Roads *et al.* (1994) point out, it is the instantaneous distribution of the moisture transport that dictates whether precipitation occurs. Moreover, summertime rainfall over the Sonoran Desert shows considerable temporal variability (Bryson and Lowry 1955; Carleton 1986; Watson *et al.* 1994), a clear indication of the importance of transience in modulating precipitation.

The distribution of the transient flux exhibits a pattern that may be related to the underlying orography. The flux vectors fan out from the Sierra Madre Occidental, indicating a transport of moisture away from the mountains. This pattern of divergent vectors appears in every season, suggesting an *intraseasonal* oscillation that is geographically fixed to the mountains and dominates the transient moisture transport. The transport away from mountains may also be related to the strength of the convection along the Sierra Madre Occidental.

While the flux for the SFC-200 mb column depicts the total horizontal transport

of water vapor, further insight is offered by examining $\bar{Q}$ above 700 mb (700-200 mb) and below 700 mb (SFC-700 mb) in conjunction with the vertical flux between the two layers. The flux above 700 mb (Fig. 7a) can be characterized as a large-scale rotation about the subtropical high, yielding easterly transport over the tropics and West Mexico and southerly transport over Arizona. The upper-level moisture over the Sonoran Desert appears to come primarily from above the Gulf of Mexico. Examination of the mean vertical flux through the 700 mb level (Fig. 8), however, indicates a major injection of moisture into the 700-200 mb layer over the Sierra Madre Occidental and the Mogollon Rim of Arizona. The band of strong upward flux coincides with the region of maximum July-August rainfall, and undoubtedly reflects the persistent convection that occurs along the mountains during the summer (Douglas *et al.* 1993; Negri *et al.* 1994; Watson *et al.* 1994). A more gradual and widespread downward transport flanks the band of strong upward transport.

The moisture flux in the SFC-700 mb layer (Fig. 7b) is not as simply characterized as that aloft. The dominant signature is the strong flux from the western Gulf of Mexico into Texas. Maximum flux vectors in the region are 2 to 3 times larger than maxima elsewhere in the figure. Flux vectors over the Mexican Plateau are generally easterly but small; this indicates little transport of moisture across the continent at low-levels. SW Arizona and NW Sonora display a strong onshore transport of moisture at low-levels from the northern Gulf of California. The flux remains onshore along the entire coastline of Sonora, but is much stronger to the north. The ECMWF analyses also suggest a noticeable low-level transport across the mountains of Baja California

Norte from the extratropical East Pacific Ocean into the northern Gulf of California. Another region of noteworthy onshore flux is into Sinaloa, Mexico where the eastern portion of a southerly stream of moisture from the tropical East Pacific terminates. Over the central Gulf of California, the flux vectors are small. Thus, the ECMWF analyses display little evidence of a time-mean transport of moisture at low-levels from the tropical Pacific into the Sonoran Desert.

Although there is little evidence of the mean flow transporting moisture into the Desert Southwest from the tropical East Pacific at low-levels, consideration of the vertical flux and the 700-200 mb horizontal transports indicates that moisture from the tropical East Pacific might reach the region aloft. The upward flux over Sinaloa (Fig. 8) would inject East Pacific moisture into the upper atmosphere where it would then cross the Gulf of California and turn northward over the Baja Peninsula. This path would carry the moisture over extreme NW Sonora.

The general features indicated by the horizontal and vertical transports qualitatively agree with the previous findings (e.g. Reyes and Cadet 1988; Roads *et al.* 1994). Quantitatively, the magnitudes of all fluxes are approximately one order of magnitude larger than our lower-bound estimate of analysis uncertainty assuming grid point accuracies comparable to radiosonde observations.

Atmospheric Balance of Water Vapor

The time-averaged balance eq.(4) for water vapor indicates that only the divergence component of the vapor flux contributes directly to the hydrological cycle. It

is difficult to infer from the horizontal flux vectors the source and sink regions of water vapor because the rotational component dominates the divergent component in many areas. This situation is clearly the case for Figs. 5 and 6a, both of which are marked by large-scale, anticyclonic curvature about the subtropical highs. For this reason, the distribution of the flux divergence of vertically-integrated water vapor was calculated.

The flux divergence for the SFC-200 mb layer (Fig. 9) reveals that, on average, sink regions are situated over the land and source regions are over the adjacent oceans. The regional distribution of sources and sinks, however, is rather complex. The largest values of convergence are situated along the western slopes of the Sierra Madre Occidental where P exceeds E by as much as 35-40 cm per month. The band of convergence extends northward along the Mogollon Rim of Arizona and southward into the tropical East Pacific. Its position coincides closely with observed precipitation maxima, and the residual P values (neglecting E over the land, a crude assumption) are consistent with, but slightly greater than, the July-August monthly means for the region (Douglas *et al.* 1993; Negri *et al.* 1994). Another region of strong convergence is located along the coastline of Texas and eastern Mexico. A major source region lies off the west coast of the Baja peninsula, with maximum values on the order of ~ 30 cm per month. This same region was found to be the strongest summertime source region for North America by Roads *et al.* (1994). The Gulf of California also shows up as a minor source region, with $E - P$ values in the range of 5-15 cm per month.

While the aforementioned source and sink regions seem both physically and quantitatively reasonable, other features of the flux divergence field are more difficult

to comprehend. In particular, the strongest source region in our analysis domain, with $E - P$ values greater than 40 cm per month, lies over the southern Rio Grande River Valley. This same feature pervades the flux divergence of all 8 summers in our data sample. The sign of the divergence may be realistic as monthly precipitation in May and June runs 10-50% larger than in July and August over NE Mexico and S Texas (Douglas *et al.* 1993; Climatological Data, Texas 1989). The greater rainfall in late spring provides the potential for evaporation to exceed precipitation during the summer. It is interesting to note that prior studies have also found this area to be a source of water vapor during the summer (Rasmusson 1966; Roads *et al.* 1994). The magnitude of the flux divergence, on the other hand, is very dubious. Comparison of the $\bar{q} \nabla \cdot \bar{\mathbf{V}}$ and $\bar{\mathbf{V}} \cdot \nabla \bar{q}$ fields suggests that the large magnitude results primarily from an overestimate of the divergence of the low-level wind in the region. It is not clear what factors cause the extreme divergence, but improper adjustment of the low-level wind to the model's distribution of terrain and/or use of only surface and mandatory level analyses in the vertical integrations (i.e. much coarser vertical resolution than the original ECMWF sigma-level analyses) are suspected.

Estimates of the spatially averaged rmse (assuming radiosonde accuracy at each grid point) suggest a uncertainty of $0.6 \text{ g cm}^{-2} \text{ mon}^{-1}$, but the large divergence over the lower Rio Grande Valley suggests that much larger errors exist in our analysis. Since surface runoff SR must equal $P - E$ (assuming surface and underground storage are negligible), some limited comparisons can be made with actual streamflow measurements for the SW United States. Using the streamflow data for the United States of Wallis *et*

al. (1991), we find that $E - P$ is about an order of magnitude too large over the Mogollon Rim, the southeast Texas coast, and most of New Mexico. Streamflow data also indicate a net runoff ($SR > 0$) in the lower Rio Grande Valley, implying an atmospheric sink of moisture exists there rather than a major source as depicted by our analysis. Runoff data for Mexico have not been compared, although precipitation data over western Mexico, as presented by Douglas *et al.* (1993), suggest the distribution of $E - P$ is realistic but $\sim 50\%$ too large.

On the whole, we believe that our estimates of the atmospheric balance are qualitatively realistic in terms of the sense of sign over most regions, but the mandatory-level ECMWF analyses alone, even at a T106 resolution, do not in general yield accurate quantitative estimates of the local balance of water vapor over the entire region.

Regional Balance of Water Vapor and Moisture Transport across Regional Boundaries

The first two parts of this study focused on documenting the water vapor flux and its divergence at the resolution of the T106 ECMWF analyses. Our diagnosis suggests that the mandatory-level analyses are not accurate enough to produce physically realistic estimates of highly differentiated quantities such as the flux divergence. For this reason, regionally-averaged flux divergences and fluxes across the regional boundaries are calculated with the goal of obtaining a more meaningful depiction of the primary sources and sink regions of water vapor.

Area-average divergences and lateral boundary fluxes (LBF's) were computed for five regions that were selected on the basis of the underlying terrain. These subregions

are illustrated in Figure 10 and were defined as follows: the United States east (USE) and west (USW) of the continental divide; Mexico east (MXE) and west (MXW) of the crest of the Mexican Plateau as portrayed by the T106 terrain field; and the Gulf of California-Pacific (GCP). Values for the SFC-700 mb layer are identified by the subscript L for low-level, and those for the 700-200 mb layer by the subscript U for upper-level.

The area-averaged moisture budgets for the SFC-200 mb column (Fig. 11) indicate that three land regions USW, USE and MXW are sinks for water vapor while the one ocean region GCP is a source. The flux convergence for MXW, the region of greatest July-August precipitation, is an order of magnitude larger than those for USW and USE. This distribution seems physically plausible. The MXE region, however, is also a source region since it contains the spurious divergences over the lower Rio Grande Valley. The sensitivity of the area-average divergences to modest shifts ($\pm$ one grid point) in their boundaries was tested. We find that the sense of divergence can be sensitive to such shifts with the exception of the MXW sector which always remains a region of strong convergence.

The LBF's for the SFC-200 mb layer indicate that $\sim 80\%$ of the total water vapor transported into the entire domain comes from over the Gulf of Mexico. All moisture from the Gulf of Mexico initially enters the MXE sector. More than half ($\sim 60\%$) of the moisture that leaves MXE heads northward into USE. The remainder ($\sim 40\%$) crosses the Mexican Plateau and enters the MXW region. Thus, $\sim 90\%$ of the moisture transported into both the USE and MXW regions comes from over the Gulf of Mexico. The majority of the moisture exiting the USE region flows northward into Colorado

and Kansas. Most of the water vapor ($\sim 80\%$) that exits the MXW region continues westward into GCP; the remainder enters the USW through SE Arizona. A comparable amount of moisture flows into the USW from the west, but most moisture entering the USW sector comes from the GCP region.

If the atmosphere above and below 700 mb are examined separately (Fig. 12), a different perspective is obtained. The sense of the LBF's for the SFC-700 mb and the 700-200 mb layers are the same as those for the SFC-200 mb layer with one important exception: the low-level flux is oriented from the GCP_L into the MXW_L . This means that moisture from the Gulf of Mexico that crosses the MXE_L - MXW_L border does not flow through the western boundary of MXW_L . Thus any moisture flowing into USW_L from the GCP_L sector must be of Pacific origin. The southerly flux across the MXW_L - USW_L border would be a mixture of water vapor from the Gulf of California, the tropical East Pacific and the Gulf of Mexico.

With inward fluxes along three of its four boundaries, strong convergence is forced in the MXW_L sector which leads to a vigorous upward transport of water vapor into MXW_U . In fact, the vertical flux from the MXW_L is the largest source of moisture for the 700-200 mb layer. In conjunction with the weaker upward fluxes over the other land sectors, vertical transports account for more than $\sim 50\%$ of the water vapor input aloft.

The sensitivity of the boundary fluxes to modest shifts ($\pm$ one grid point) in the position of the regional boundaries was also examined. We find that the magnitudes of the boundary fluxes typically change by 10%, but their components normal to the boundaries remain the same with the exception of the weak 700-200 mb flux along the

western boundary of USW and the 700 mb vertical flux over GCP. The impact of raising the layer interface from 700 mb to 600 mb was also examined and found to not affect the sense of the fluxes. Thus we believe the results concerning the strong boundary fluxes discussed above are qualitatively robust.

Discussion and Conclusions

In this study, the origins and transport of water vapor into the semi-arid Sonoran Desert region of North America were examined for the July-August monsoon season. Mandatory-level analyses produced by ECMWF at a spectral truncation of triangular 106 were used to compute vertically-integrated fluxes and flux divergences of water vapor.

The T106 ECMWF analyses, interpolated to only the mandatory-levels, do not possess sufficient precision to yield accurate estimates of highly differentiated quantities such as the divergence of the vertically-integrated flux of water vapor. Perhaps calculation of the water balance using the T106 ECMWF analyses on their original sigma layers would remedy the problem of dubious source regions over the North American continent (such as the major source over the lower Rio Grande River Valley); calculations on sigma levels have proven useful in prior studies that used model output to compute budgets for other atmospheric quantities (e.g. Sardeshmukh and Held, 1984; Mullen, 1986). In any event, we believe that the T106 mandatory-level analyses, for the most part, provide a faithful portrait of the vertically-integrated, horizontal transport of water vapor.

As a way to summarize the most robust results of this study, we offer the schematic diagram of Fig. 13. The figure portrays the primary streams of water vapor by the time-mean circulation for July-August as depicted by the T106 ECMWF analyses. These results can be summarized as follows:

- The transport of water vapor by the time-mean circulation dominates that due to the transient eddies.
- Most water vapor enters the Sonoran region at low-levels (below 700 mb). This moisture comes primarily from over the northern Gulf of California, but limited amounts from the Gulf of Mexico flow over the Sonoran Gap into the eastern Sonoran Desert. The time-mean flow is unable to transport water vapor directly from the southern Gulf of California and/or tropical East Pacific to the Sonoran Desert at low-levels.
- Most of the upper-level (above 700 mb) moisture over the Sonoran Desert appears to come from over the Gulf of Mexico, circulating around the southern and western quadrants of the the subtropical ridge.
- Onshore, low-level flow from the southern Gulf of California and the tropical East Pacific produces a convergence of water vapor that helps fuel the persistent convection along the west slopes of the Sierra Madre Occidental. The convection injects a large amount of water vapor into the upper-levels. Once aloft, this tropical moisture may reach the western Sonoran Desert after subsequent northwestward transport by the mean flow.

While our results support the notion that the northern Gulf of California is a major source of low-level moisture for the Sonoran Desert, evidence exists that the impact of the Gulf on the regional climate is not properly resolved by the T106 ECMWF analyses. The land-sea mask for the ECMWF analyses (Fig. 14) reveals that the northern Gulf of California is represented by two inland seas at a T106 truncation. The ECMWF terrain field (Fig. 14) indicates that mountains of the Baja Peninsula are also poorly resolved, especially over Baja California Norte where the actual elevations along the crest run higher than 1000 m but the analysis system produces heights around 500 m. The lower barrier in the T106 ECMWF analysis system could prevent the blocking of the low-level northwesterlies to the west of Baja California Norte and lead to an excessive penetration of low-level westerly momentum across the peninsula into the northern Gulf of California. This could help explain surface southwesterlies in the T106 ECMWF analyses over a region where observations from special field programs (Badan-Dangon *et al.* 1991; Douglas *et al.* 1993) indicate prevailing S-SSE winds exist.

The inadequate resolution of the local geography at a T106 truncation has important consequences for future studies that examine the North American monsoon. For example, current reanalysis efforts will not address the situation because they employ horizontal resolutions of T106 or coarser (e.g. Kalnay and Jenne, 1991; Gibson *et al.* 1994; Janowiak *et al.* 1994; Kistler *et al.* 1994). Thus we believe that diagnostic studies based on the reanalysis output may provide little additional insight into the impact of the Gulf of California and regional-scale circulations on the monsoon over that already obtainable from the current T106 ECMWF analyses.

Clearly resolutions finer than T106 are required before the regional-scale circulations associated with the North American monsoon can be understood. We believe that simulations with regional climate models (RCMs), employing horizontal and vertical resolutions higher than the T106 ECMWF mandatory-level analyses, offer the most economical way to achieve this goal. However, the ability of RCMs to faithfully simulate the monsoon has not been firmly established. Pioneering simulations (e.g. Giorgi 1991; Giorgi *et al.* 1994) of the Sonoran region missed the most important features of the monsoon such as maximum summer precipitation being over western Mexico. Using four-dimensional data assimilation to ingest special field observations, Stensrud *et al.* (1995) showed that 24 h simulations with a suitably constructed, properly initialized mesoscale model can capture many of the salient features of the monsoon, such as low-level mean flow up the Gulf of California and the axis of maximum precipitation along the western foothills of the Sierra Madre Occidental. While the results of Stensrud *et al.* (1995) are encouraging, they did not examine whether simulations would be nearly as accurate if the special observations were excluded or if the integrations were extended beyond 24 h. Ultimately, long-term observations in the immediate vicinity of the Gulf of California, similar in scope to those analyzed by Badan-Dangon (1991) and Douglas *et al.* (1993) for single summers, will be needed to validate climate models, to use for mesoscale data assimilations, and to determine the true role of the Gulf of California.

Even though our results indicate that the time-mean circulation dominates the transport of water vapor, we believe that the role of transience needs to be thoroughly

examined. As previously noted, the temporal variability in summertime rainfall over the Sonoran region attests to the importance of transience in modulating precipitation (Bryson and Lowry 1955; Carleton 1986; Watson *et al.* 1994). Research is underway using the T106 ECMWF analyses that contrasts the larger-scale circulation and moisture transport during “bursts” and “breaks” in the monsoonal rainfall over the Sonoran Desert. Results will be reported in due course.

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Appendix 1.

The water vapor balance per unit mass for a parcel of air as given by Rasmusson (1977) is

$$\frac{dq}{dt} = \frac{\partial q}{\partial t} + \vec{V} \cdot \nabla q + \omega \frac{\partial q}{\partial p} = e - c + g \frac{\partial D}{\partial p}, \quad (A1)$$

where g = gravitational acceleration,

t = time,

p = pressure,

$\vec{V}$ = horizontal velocity vector,

ω = pressure vertical velocity,

e = evaporation per unit mass for an air parcel,

c = condensation rate per unit mass for an air parcel,

q = specific humidity, and

D = vertical diffusion rate per unit area of atmospheric water vapor.

With the aid of the mass continuity equation (A1) can be transformed into

$$e - c + g \frac{\partial D}{\partial p} = \frac{\partial q}{\partial t} + \nabla \cdot q \vec{V} + \frac{\partial \omega q}{\partial p} \quad (A2)$$

which is the water vapor balance equation for a particular isobaric level. The total transport within a column of air is obtained by integrating with respect to pressure from

the top of the atmosphere p_{top} to the surface of the earth p_{sfc} which gives

$$\int_{p_{\text{top}}}^{p_{\text{sfc}}} (e - c) \frac{dp}{g} + \int_{p_{\text{top}}}^{p_{\text{sfc}}} g \frac{\partial D}{\partial p} \frac{dp}{g} = \int_{p_{\text{top}}}^{p_{\text{sfc}}} \frac{\partial q}{\partial t} \frac{dp}{g} + \int_{p_{\text{top}}}^{p_{\text{sfc}}} \nabla \cdot q \vec{V} \frac{dp}{g} + \int_{p_{\text{top}}}^{p_{\text{sfc}}} \frac{\partial \omega q}{\partial p} \frac{dp}{g} . \quad (\text{A3})$$

At this point Leibnitz's Rule (e.g. Hildebrand 1962, p 360) is employed which states given a function

$$F(x, y) = \int_{A(x)}^{B(x)} f(x, y) dy$$

the partial derivative of F with respect to x has the form

$$\frac{\partial F(x, y)}{\partial x} = \int_{A(x)}^{B(x)} \frac{\partial f(x, y)}{\partial x} dy + \left[f(x, B) \frac{\partial B}{\partial x} \right] - \left[f(x, A) \frac{\partial A}{\partial x} \right] .$$

Applying Leibnitz's Rule to the first two terms on the right hand side of (A3), we get

$$\begin{aligned} & \int_{p_{\text{top}}}^{p_{\text{sfc}}} \frac{\partial q}{\partial t} \frac{dp}{g} + \int_{p_{\text{top}}}^{p_{\text{sfc}}} \nabla \cdot q \vec{V} \frac{dp}{g} \\ &= \frac{\partial W}{\partial t} + \nabla \cdot \langle q \vec{V} \rangle - \frac{q_{\text{sfc}}}{g} \left[\frac{\partial p_{\text{sfc}}}{\partial t} + \vec{V}_{\text{sfc}} \cdot \nabla p_{\text{sfc}} \right] + \frac{q_{\text{top}}}{g} \left[\frac{\partial p_{\text{top}}}{\partial t} + \vec{V}_{\text{top}} \cdot \nabla p_{\text{top}} \right] \end{aligned}$$

where

$$\langle () \rangle = \int_{p_{\text{top}}}^{p_{\text{sfc}}} () \frac{dp}{g} ,$$

$W = \langle q \rangle$ is the precipitable water, and $\bar{Q} = \langle q \vec{V} \rangle$ is the vertically-integrated water vapor flux. The last term in brackets is zero since p_{top} is fixed at 200 mb. The first term in brackets is ω_{sfc} since

$$\omega_{\text{sfc}} = \frac{\partial p_{\text{sfc}}}{\partial t} + \vec{V}_{\text{sfc}} \cdot \nabla p_{\text{sfc}}$$

at $p = p_{\text{sfc}}$ (e.g. Panofsky 1946; Haltiner and Williams 1980; Trenberth 1991). After simplification, we finally obtain

$$\int_{p_{\text{top}}}^{p_{\text{sfc}}} \frac{\partial q}{\partial t} \frac{dp}{g} + \int_{p_{\text{top}}}^{p_{\text{sfc}}} \nabla \cdot q \vec{V} \frac{dp}{g} = \frac{\partial W}{\partial t} + \nabla \cdot \bar{Q} - \frac{q_{\text{sfc}} \omega_{\text{sfc}}}{g} \quad (\text{A4})$$

Substitution of (A4) into (A3) and integration of all terms results in

$$-P + D(p_{\text{sfc}}) - D(p_{\text{top}}) = \frac{\partial W}{\partial t} + \nabla \cdot \bar{Q} - \frac{q_{\text{top}} \omega_{\text{top}}}{g}, \quad (\text{A5})$$

where it is generally assumed (Rasmusson 1977) that

$$\int_{p_{\text{top}}}^{p_{\text{sfc}}} (e - c) \frac{dp}{g} = -P,$$

From (A5) it is clear that $\omega_{\text{sfc}} q_{\text{sfc}}/g$ produced by vertically integrating the vertical flux divergence terms cancels the $\omega_{\text{sfc}} q_{\text{sfc}}/g$ resulting from application of Leibnitz's Rule. Historically, this term has been treated in one of two ways: most commonly it is set to zero by assuming ω_{sfc} is zero (Palmén 1967; Rasmusson 1967,

1968, and 1977)–i.e. a “flat earth” assumption; less frequently it is used to represent evapotranspiration at scales too small to resolve (Peixoto 1973).

Several more terms can be eliminated following the integration of (A5).

$\omega_{\text{top}} q_{\text{top}}/g$ becomes negligible since $q_{\text{top}} \approx 0$. Similarly, the vertical diffusion at the top of the atmosphere $D(p_{\text{top}})$ can be neglected. This leaves

$$E - P = \frac{\partial W}{\partial t} + \nabla \cdot \tilde{Q} , \quad (\text{A6})$$

where

$$\lim_{p \rightarrow p_{\text{sfc}}} D(p) = E .$$

Equation (A6) is the general water balance equation of the atmosphere with horizontal diffusion and the contributions from liquid and solid phases neglected.

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Figure Captions

Figure 1. Geography of the SW North American region showing locations mentioned in the text.

Figure 2. Time-mean winds and specific humidities for July-August at (a) the surface and (b) 500 mb. Maximum wind vector is 6.8 m sec^{-1} and contour interval for q is 2 g kg^{-1} in (a). Maximum wind vector is 6.3 m sec^{-1} and contour interval for q is 0.25 g kg^{-1} in (b). Vector scaling is the same for both panels. For sake of clarity, the boundaries for all figures have been cropped at 15°N , 39°N , 90°W and 120°W .

Figure 3. Time-mean vertical velocity for July-August at 500 mb. Contour interval is $0.5 \text{ microbars sec}^{-1}$.

Figure 4. Time-mean precipitable water for July-August. Contour interval is 0.5 g cm^{-2} .

Figure 5. July-August, total vector flux of water vapor for the surface to 200 mb layer. Maximum vector is $29.6 \times 10^2 \text{ g cm}^{-1} \text{ sec}^{-1}$.

Figure 6. July-August, vector flux of water vapor due to the time-mean wind for the surface to 200 mb layer due to (a) the time-mean flow and (b) the transient eddies. Maximum vector is $29.9 \times 10^2 \text{ g cm}^{-1} \text{ sec}^{-1}$ in (a). Vectors in (a) are scaled the same as those in Fig. 5. Maximum vector is $2.6 \times 10^2 \text{ g cm}^{-1} \text{ sec}^{-1}$ in (b). Vectors in (b) are magnified by a factor of five for visual clarity.

Figure 7. July-August, total vector flux of water vapor for (a) the 700 mb to 200 mb layer and (b) for the surface to 700 mb layer. Vectors in (a) and (b) are scaled the

same as those in Figs. 5 and 6a. Maximum vector is $8.3 \times 10^2 \text{ g cm}^{-1} \text{ sec}^{-1}$ in (a). Maximum vector is $23.6 \times 10^2 \text{ g cm}^{-1} \text{ sec}^{-1}$ in (b).

Figure 8. July-August, total vertical flux of water vapor through the 700 mb layer.

Contours are every $\pm 5, \pm 15, \pm 25, \dots \text{ g cm}^{-2} \text{ mon}^{-1}$; dashed lines denote upward flux.

Figure 9. July-August, flux divergence of the total water vapor transport for the surface to 200 mb layer. Contours are every $\pm 5, \pm 15, \pm 25, \dots \text{ g cm}^{-2} \text{ mon}^{-1}$; dashed contours denote convergence.

Figure 10. Geographical areas used in the calculations of the regional water vapor balances and moisture transport across regional boundaries.

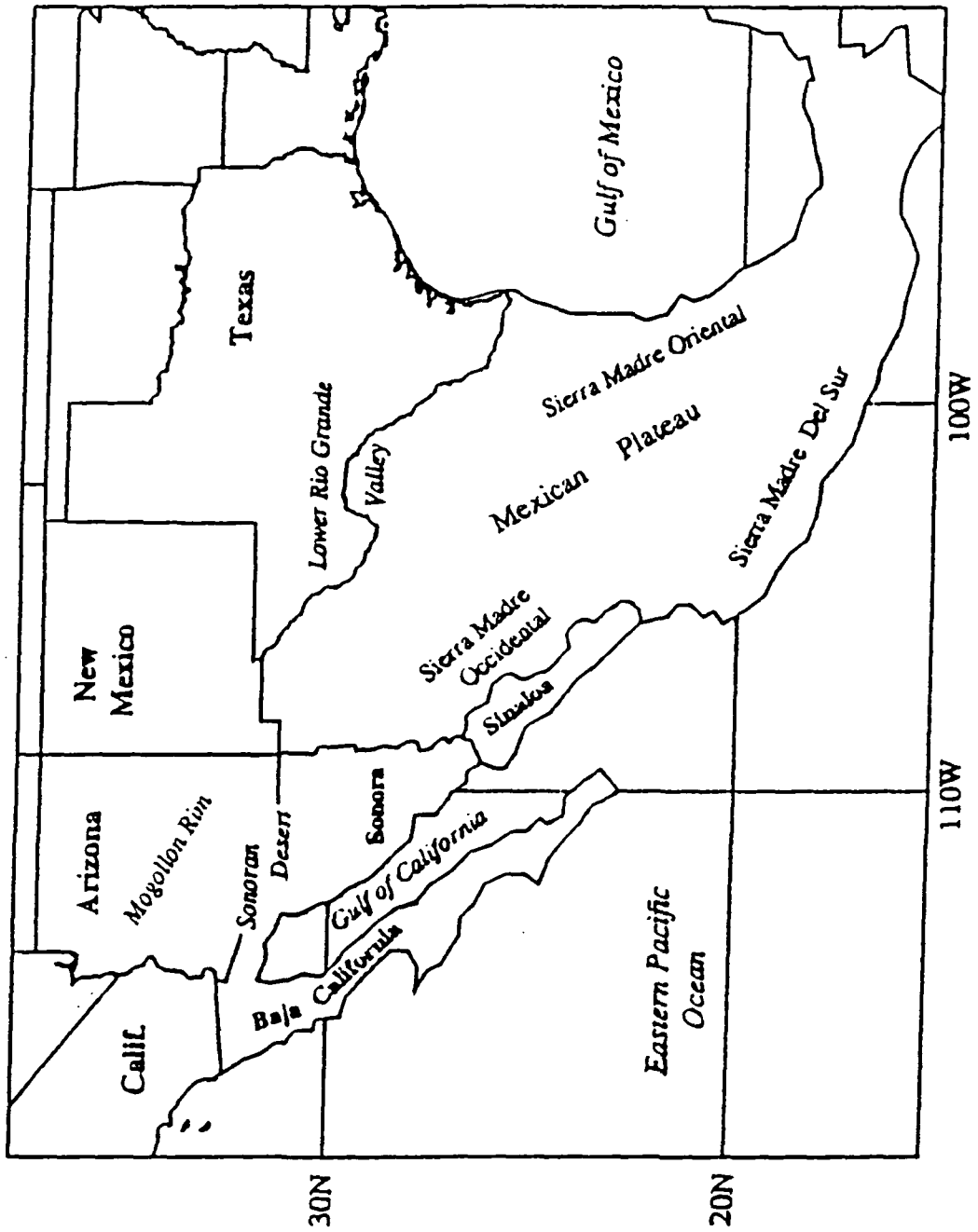
Figure 11. July-August, regional water vapor balance and horizontal fluxes across regional boundaries for the surface to 200 mb layer. Dark arrows indicate the direction of the net horizontal transport across each face of the volume. Magnitudes are presented as 10^{16} grams per month. Numbers in parenthesis indicate the net flux divergence (positive number) for the adjacent volume. Estimated uncertainties for all quantities are indicated by the $\pm$ number following each value. See text for further details.

Figure 12. July-August, horizontal fluxes of water vapor across regional boundaries for the 700 mb to 200 mb layer (top) and surface to 700 mb layer (bottom), and vertical fluxes through the 700 mb interface. Dark arrows indicate the direction of the net horizontal transport across each face of the volume; gray arrows denote the direction of net vertical transport across the 700 mb level. Plain (underlined) values

represent the (vertical) moisture transport through the 700 mb level. Magnitudes are presented as 10^{16} grams per month. Estimated uncertainties for all quantities are indicated by the $\pm$ number following each value.

Figure 13. Schematic of the most robust features of the 3-D transport of moisture based as inferred from the T106 ECMWF analyses. The cross-hatched arrows denote the primary streams of low-level (sfc-700mb) moisture; the shaded arrow denotes the primary stream of the moisture aloft (700-200mb). Width of the arrows is proportional to the magnitude of the horizontal moisture flux. Cumulonimbus clouds denote the region of strongest upward flux of water vapor and maximum precipitation.

Figure 14. Terrain heights (contour interval 500 m, with zero contour omitted) and the land/sea mask (hatching indicates ocean areas) for the T106 ECMWF analyses.



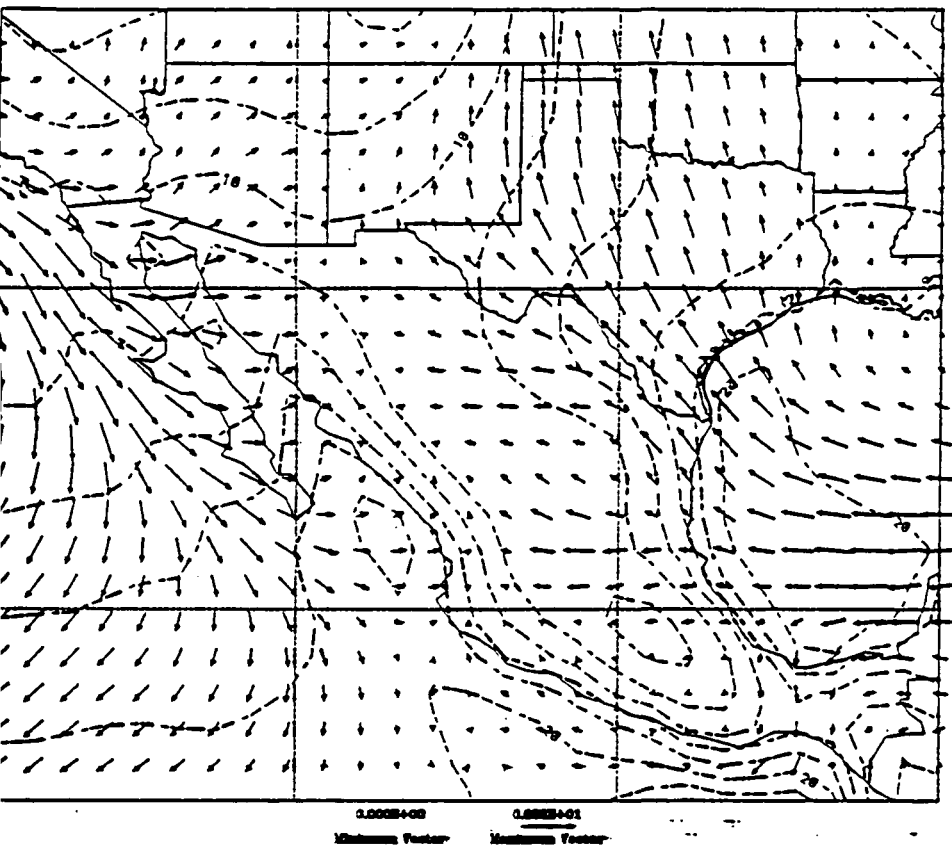


Fig. 2a

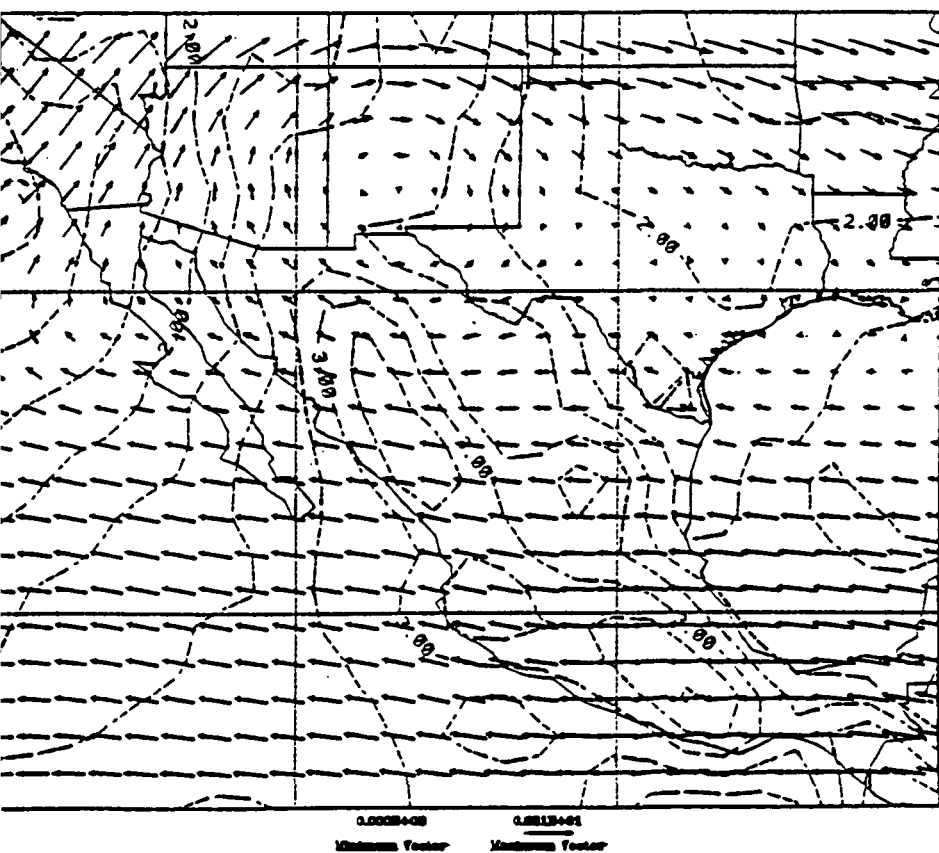


Fig. 2b

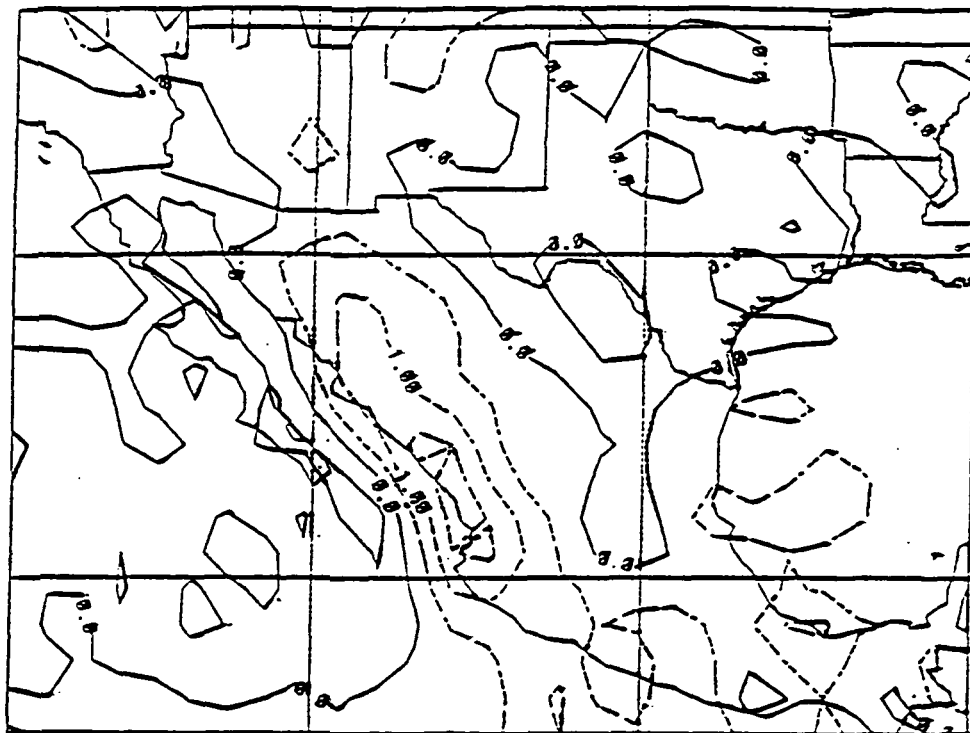


Fig. 3

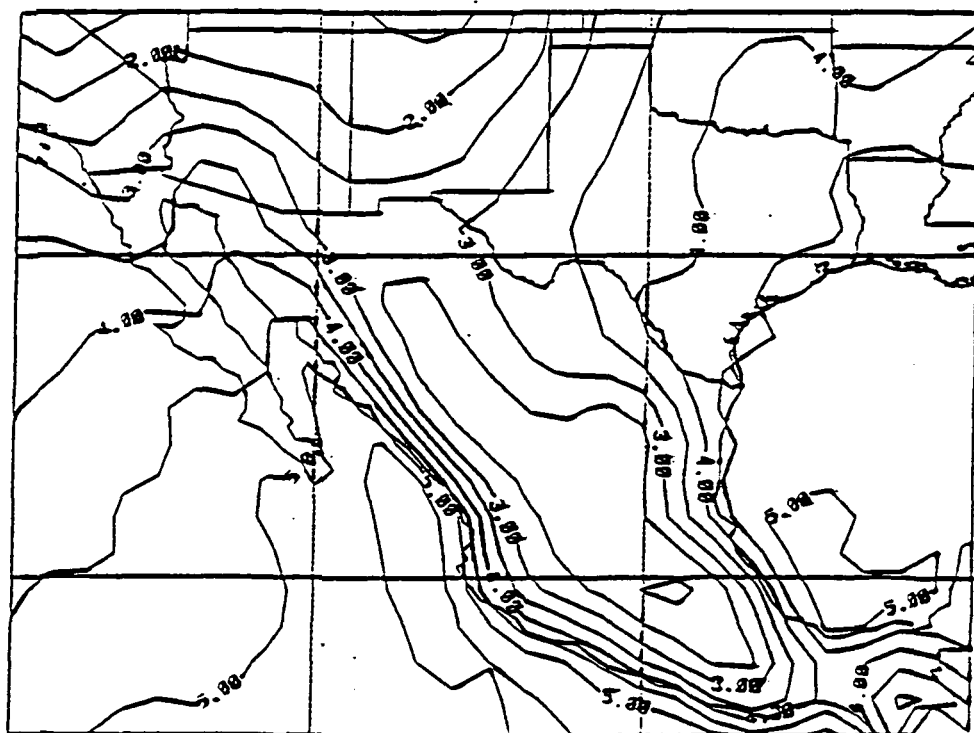
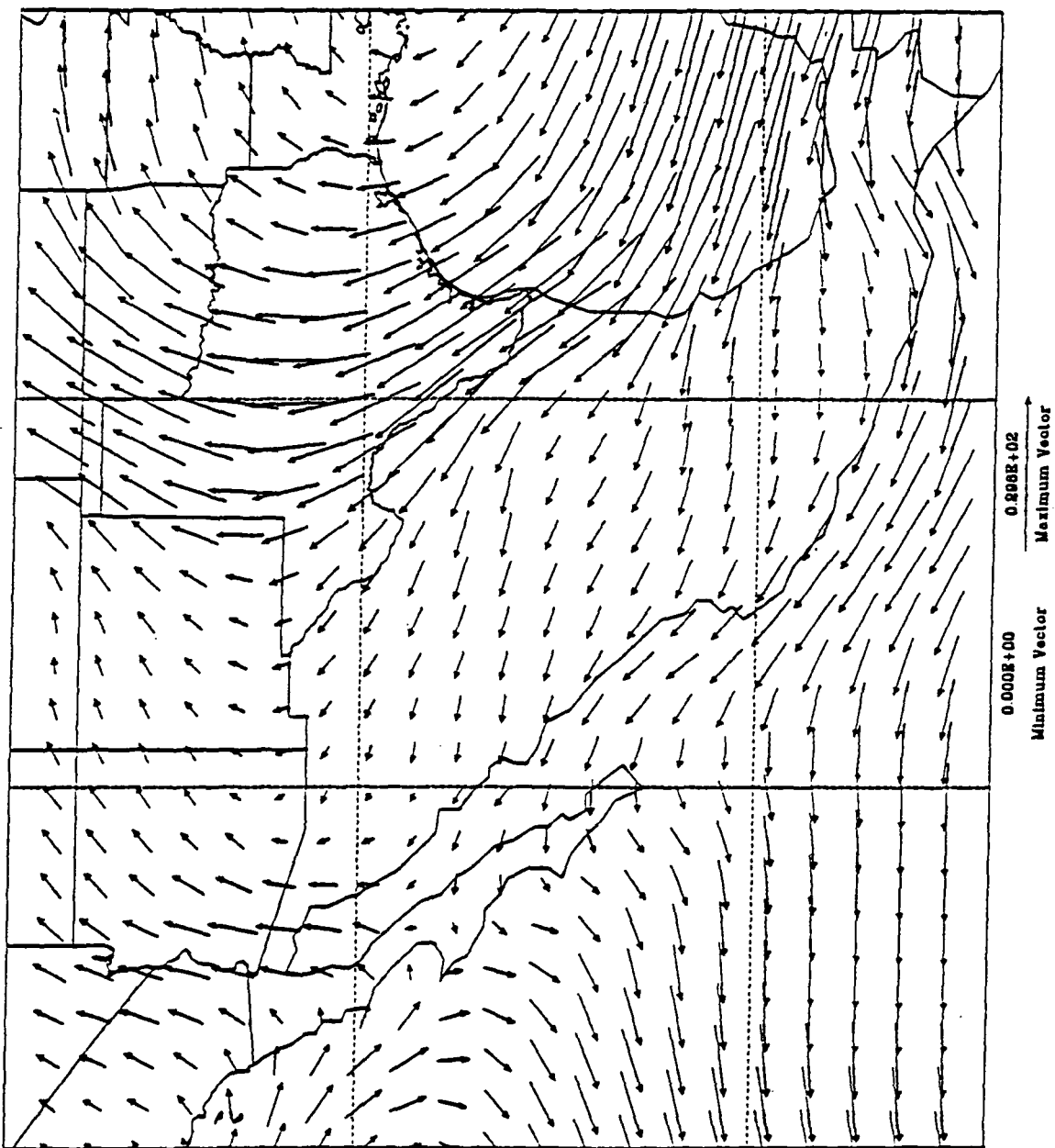


Fig. 4.

Fig. 5



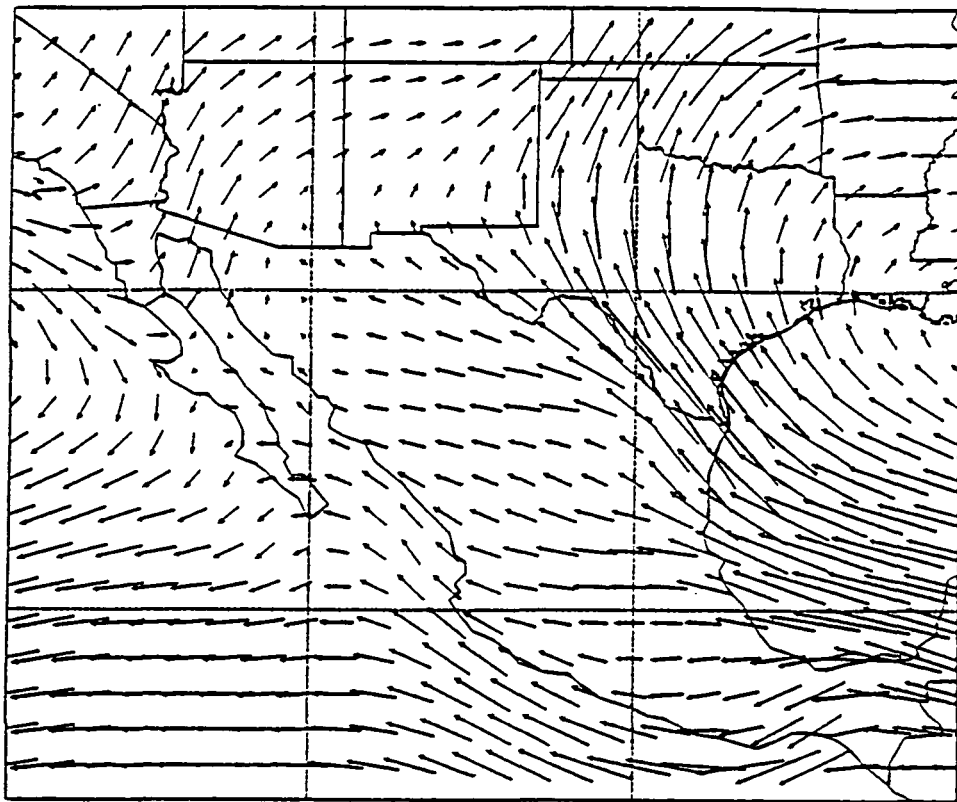


Fig. 6a

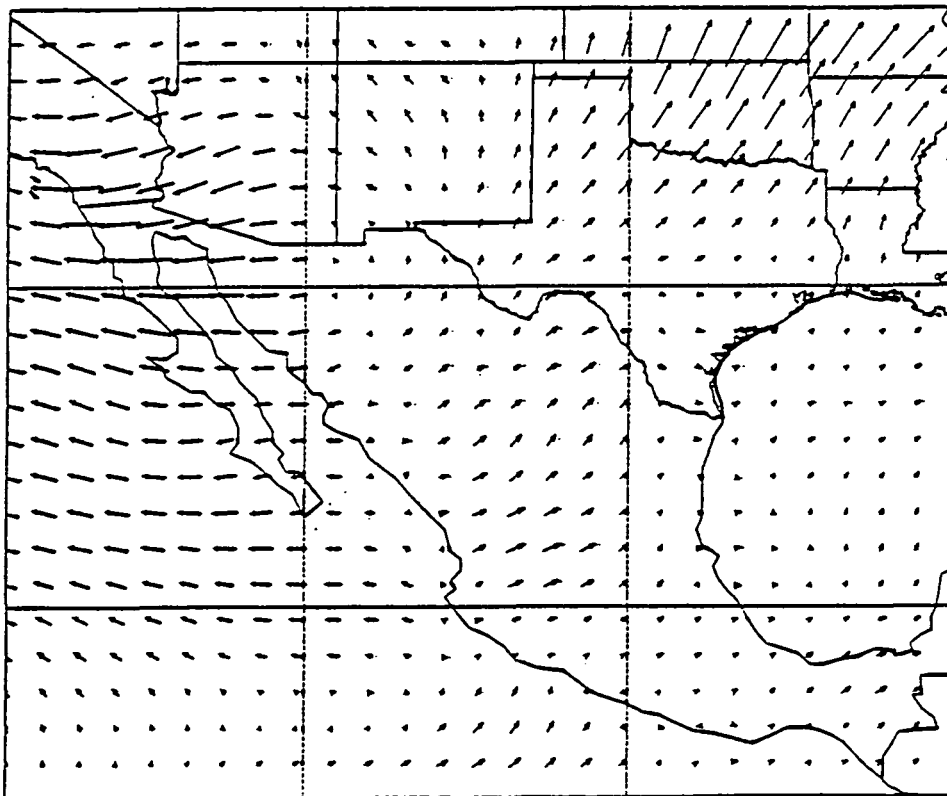
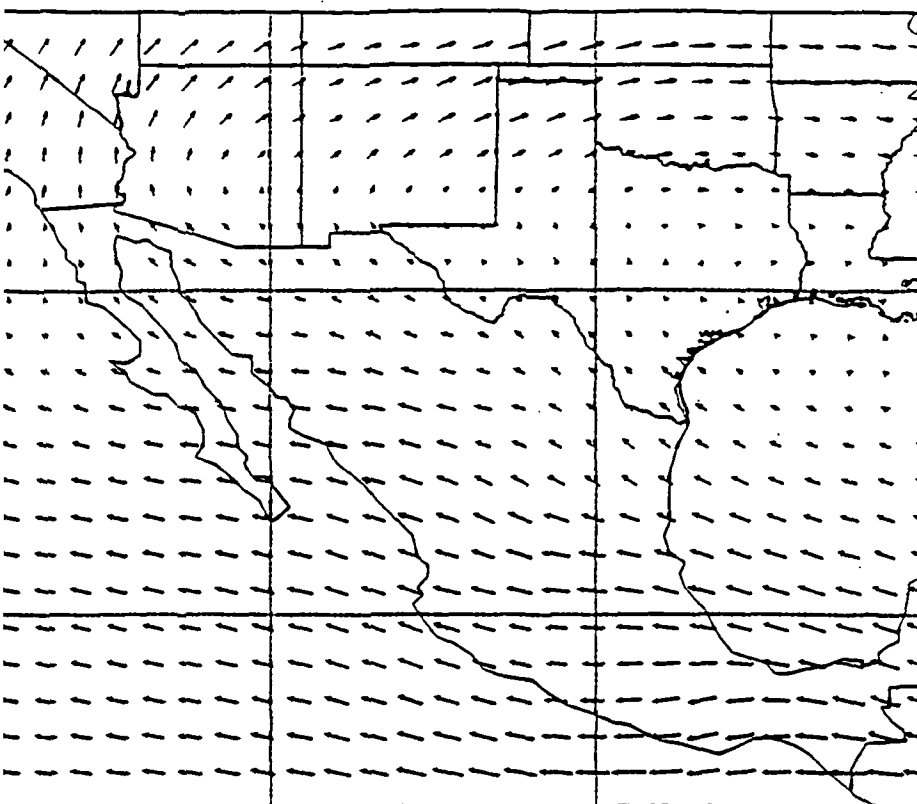
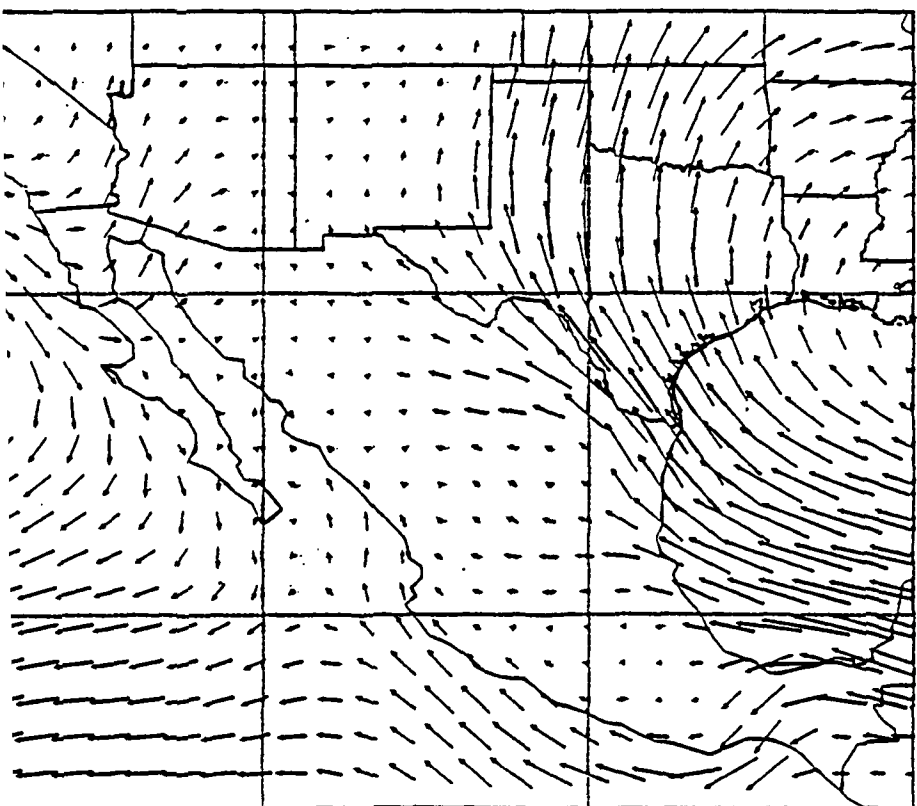


Fig. 6b



0.0002E+00 0.0002E+01
 Minimum Vector Maximum Vector

Fig 7a



0.0002E+00 0.0002E+00
 Minimum Vector Maximum Vector

Fig 7b.

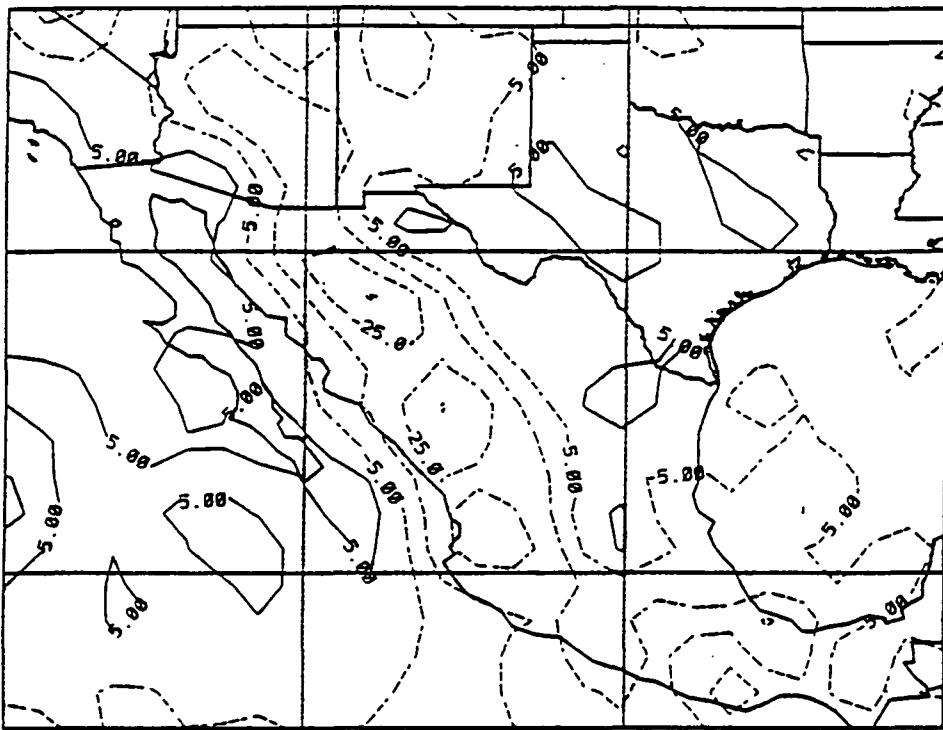


Fig 8

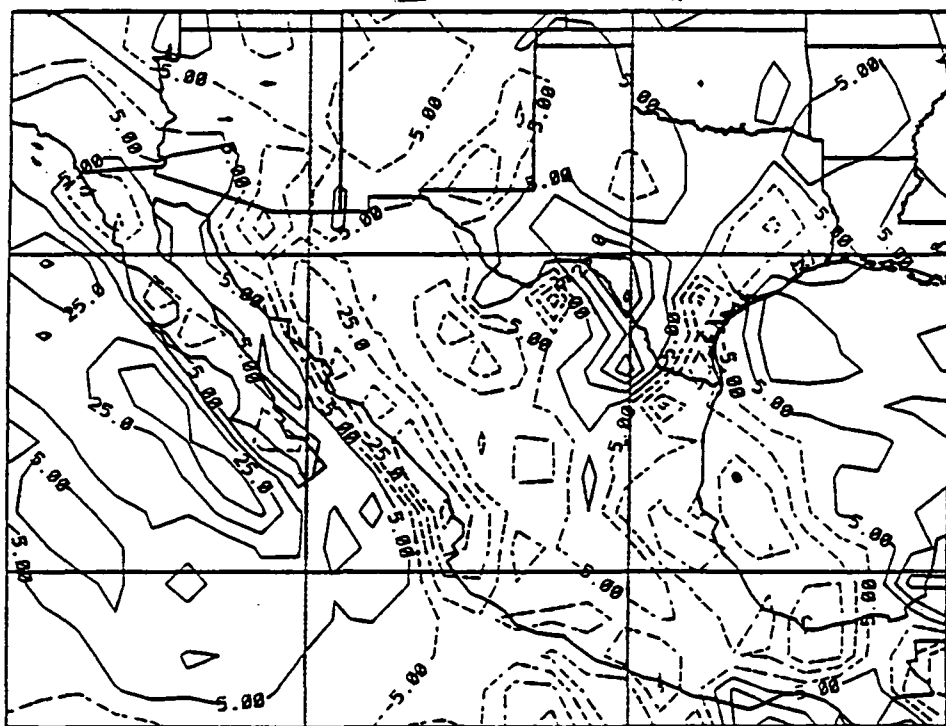
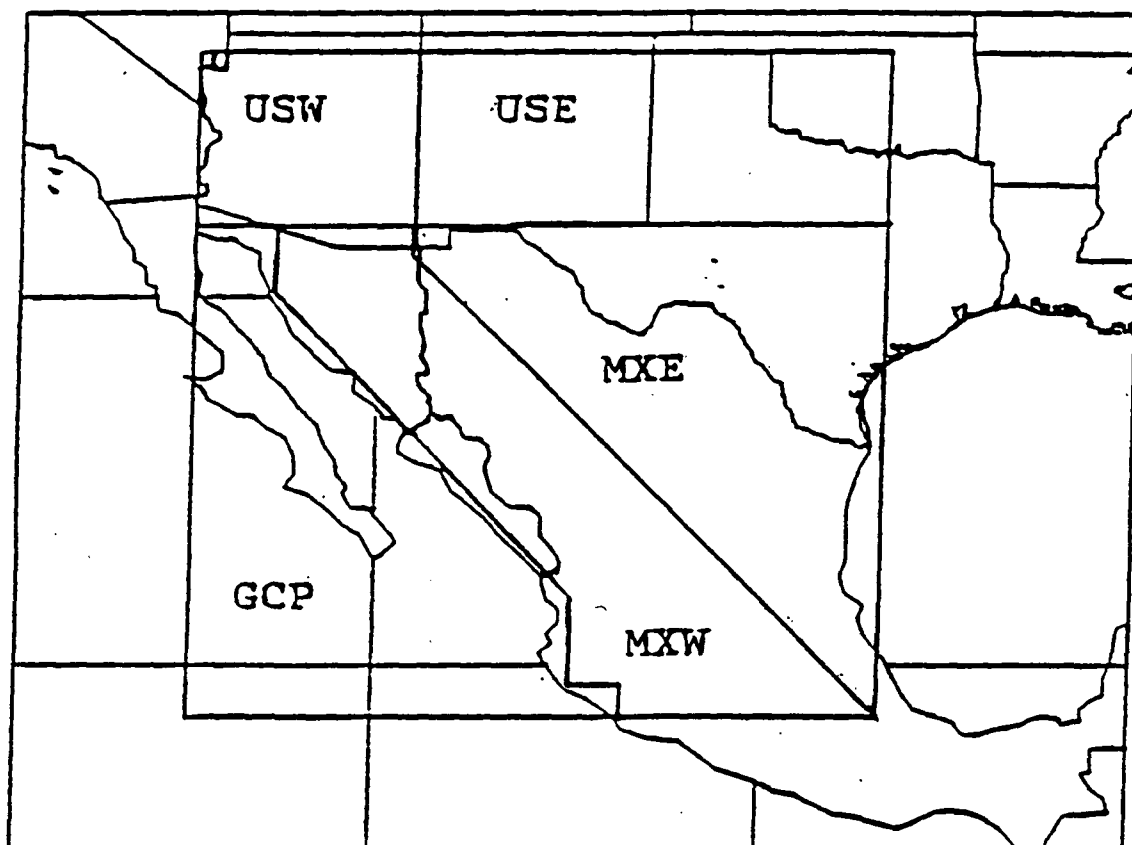


Fig 9



• Fig 10

Surface - 200mb

Daily

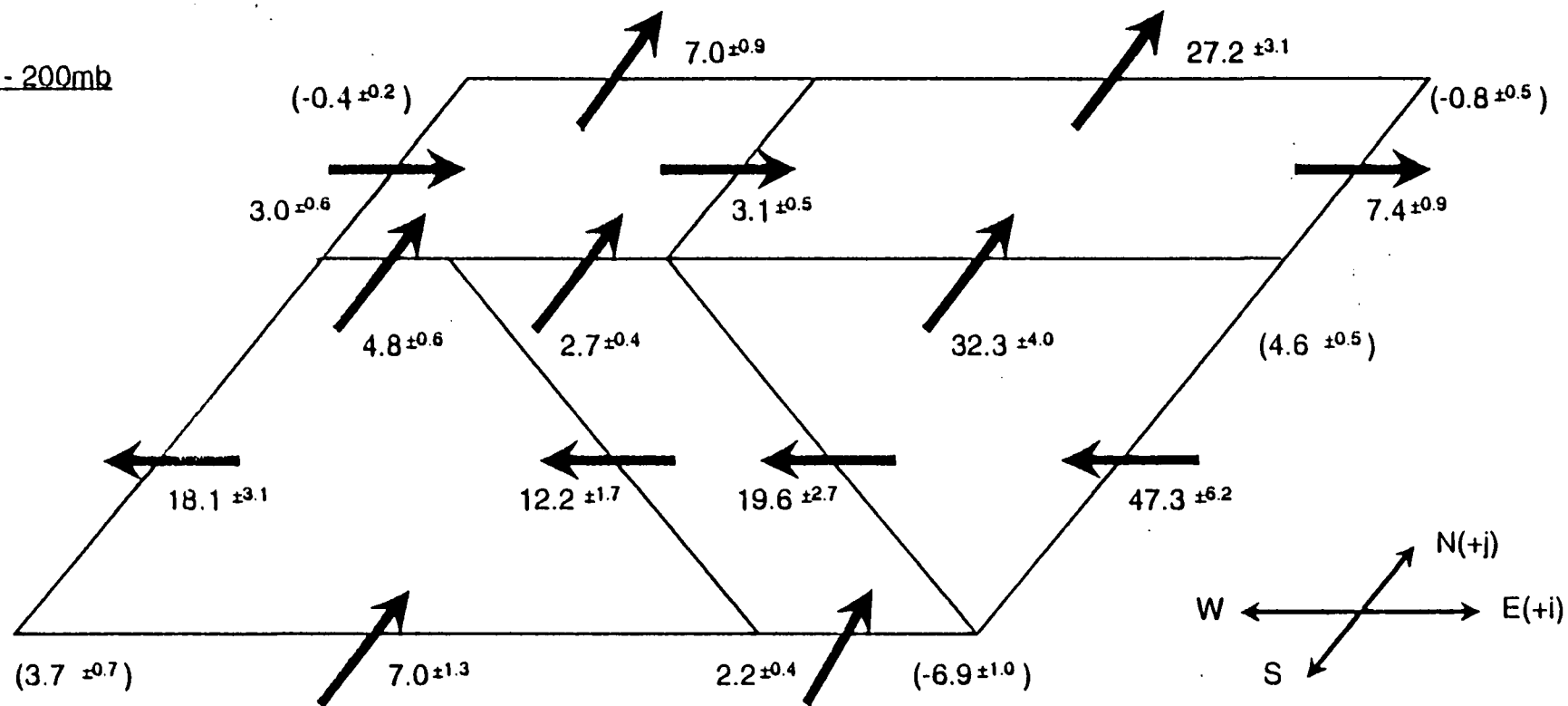
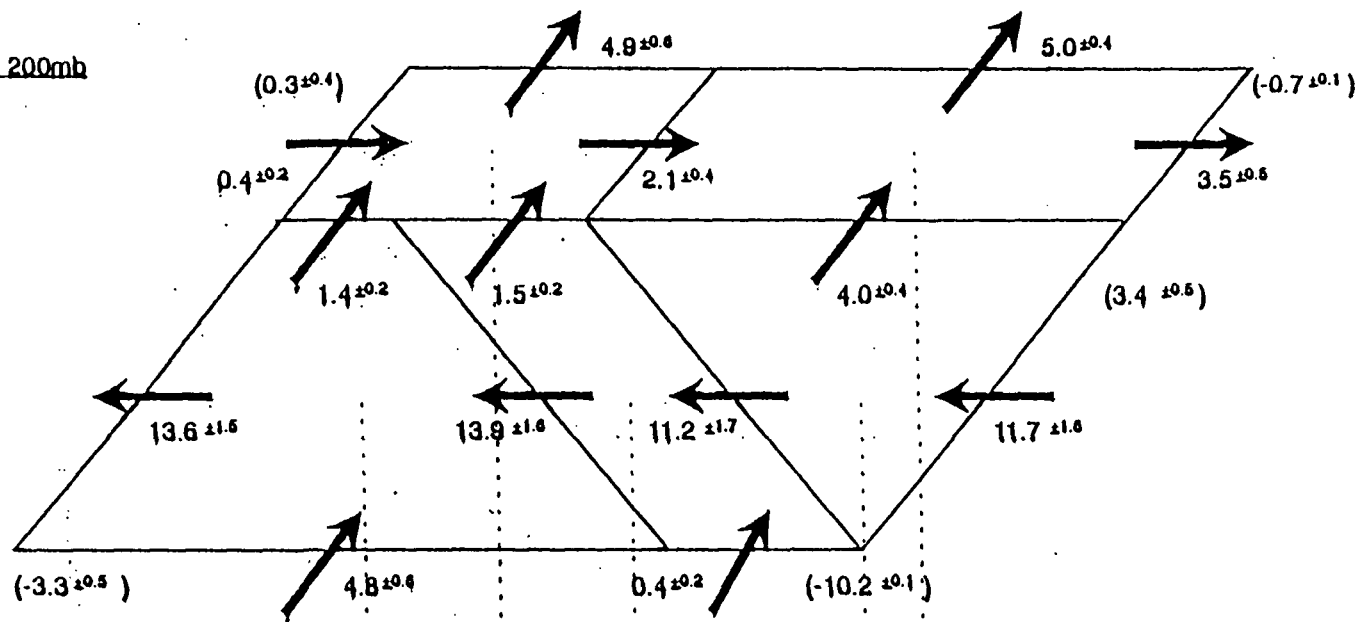


Fig 11

700mb - 200mb

Daily



Surface - 700mb

Daily

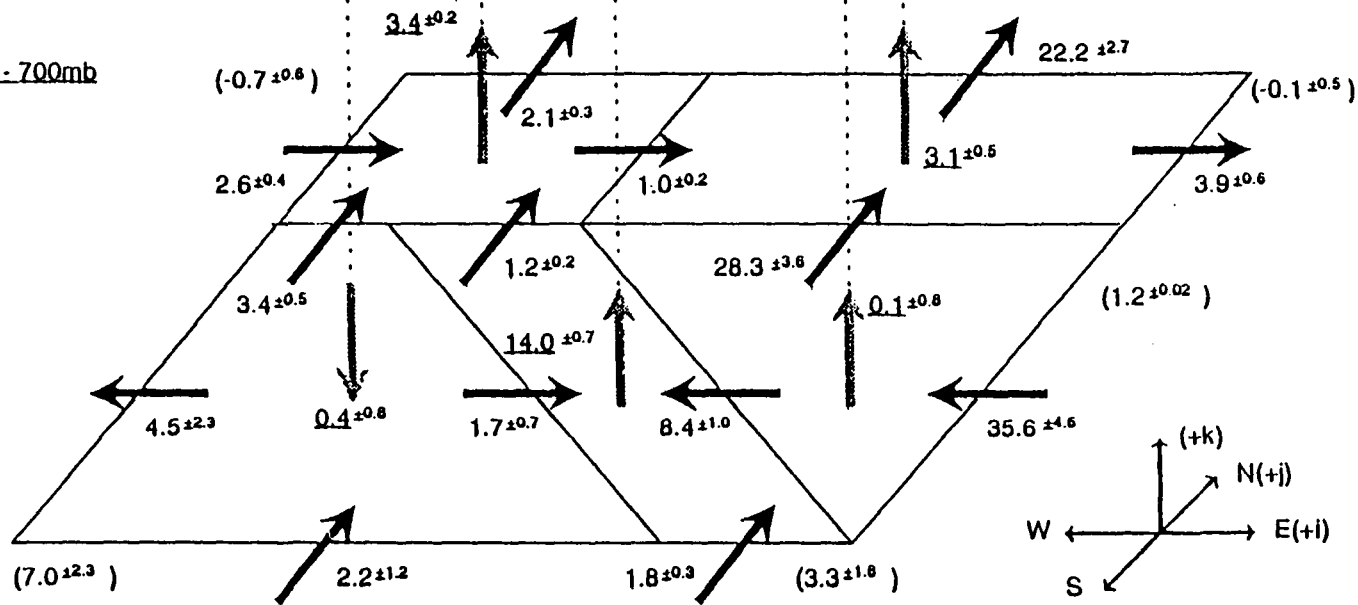


Fig 12

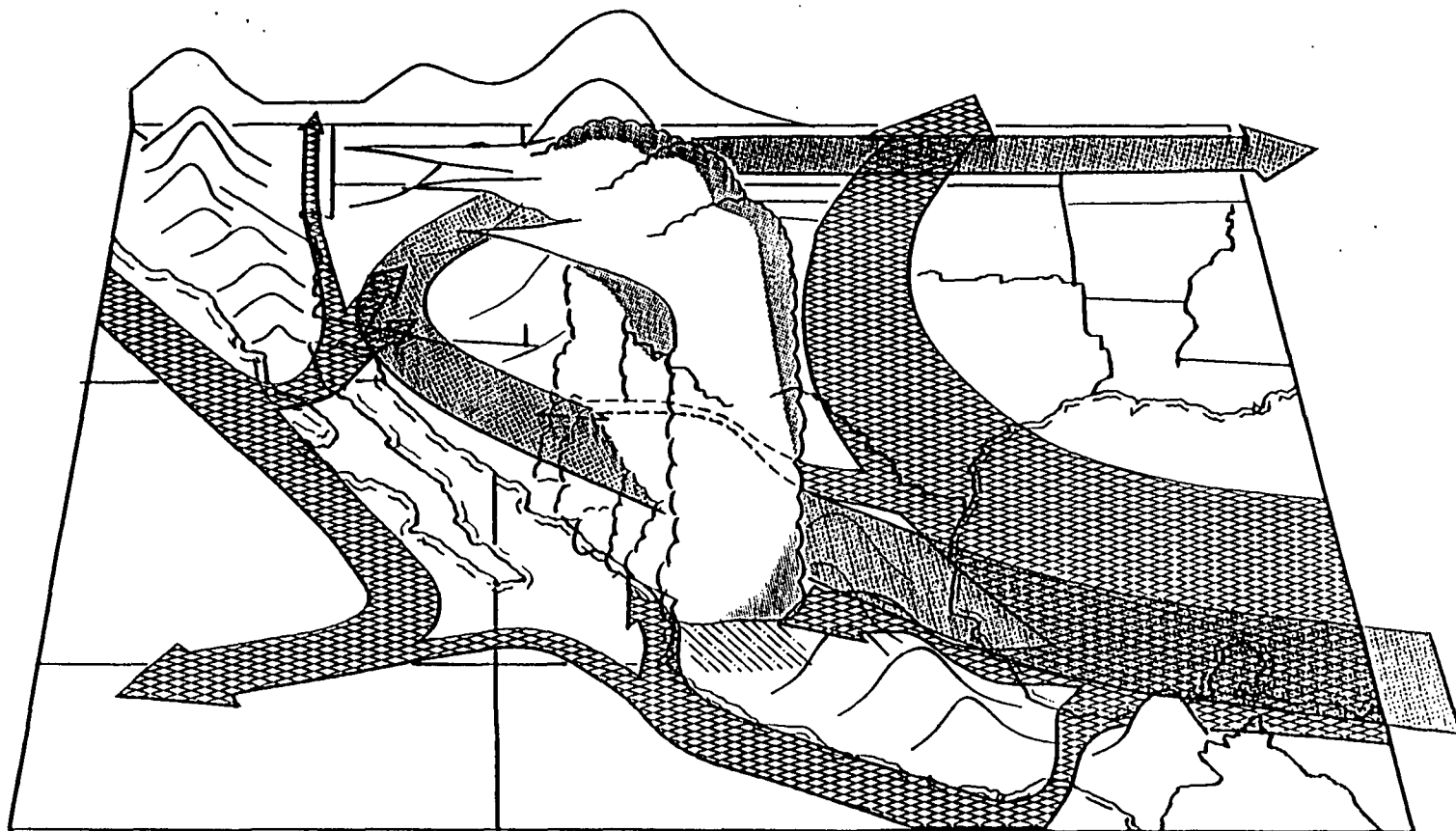


Fig 13

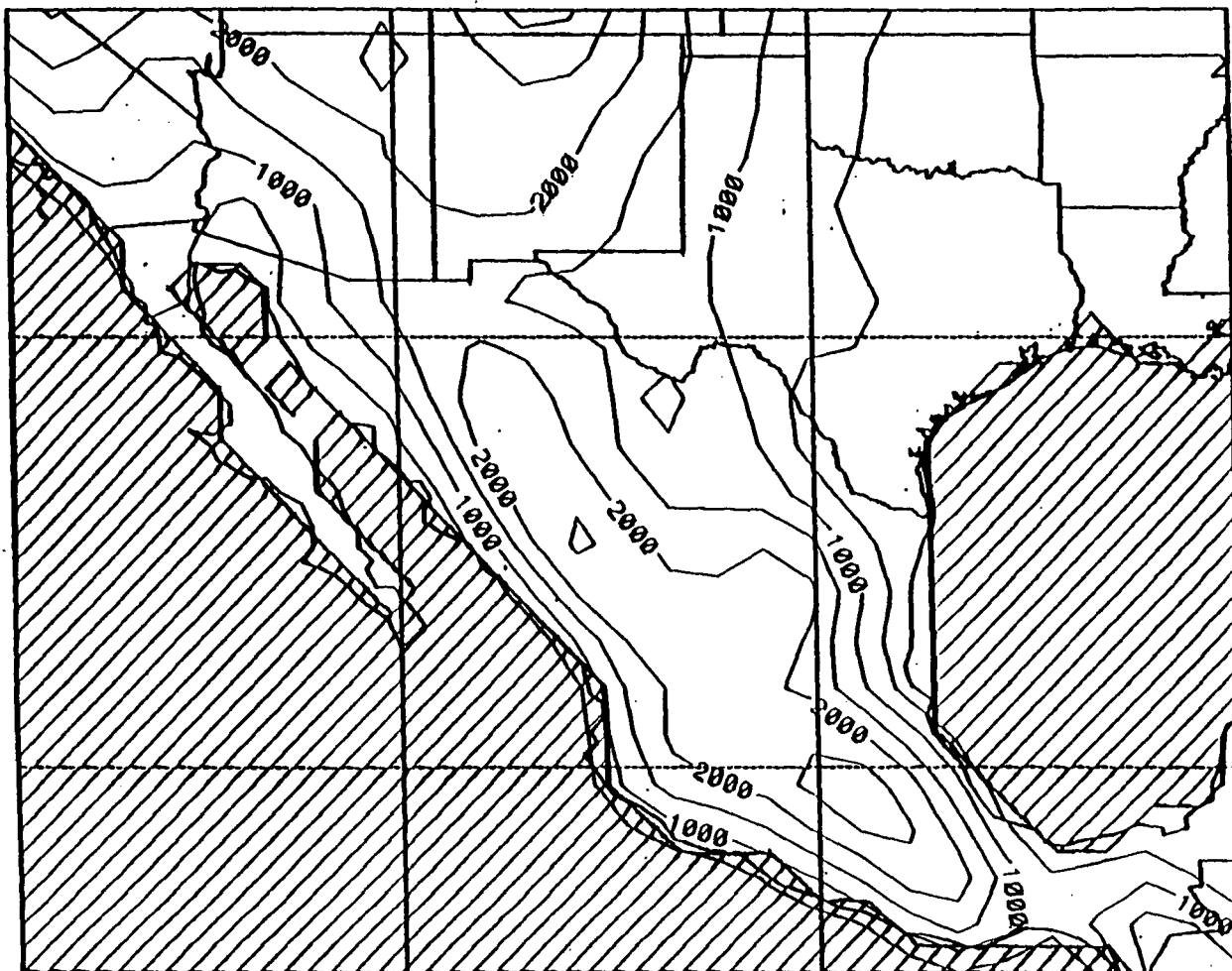


Fig. 14

APPENDIX V — Intraseasonal Variability of the Summertime North American Monsoon

**Intraseasonal Variability of the
Summertime North American Monsoon**

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1 Abstract

Intraseasonal variations associated with the North American Summer Monsoon are investigated. Composite *wet* and *dry* periods during July and August of 1985-1992, defined from rain gauge data for southeast Arizona, are compared. Cloud top temperature (CCT), horizontal and vertical velocities, specific humidity, precipitable water (PW), convective indices, moisture flux, and parcel trajectories are all examined. ECMWF mandatory-level analyses possessing a spectral resolution of triangular 106 are employed.

Significant differences exist between *wet* and *dry* conditions over the Sonoran Desert for all fields considered. As the monsoon shifts from *dry* to *wet* conditions, the subtropical ridge shifts $\sim 5^\circ$ latitude toward the north, and PW increases by as much as ~ 1.2 cm (~ 0.5 inches). Parcels in the middle troposphere ascend into the region from the southeast, and the atmosphere becomes more unstable. The result is a significant increase in the frequency of deep convection, as determined from $CTT < -38^\circ\text{C}$.

During both monsoon regimes, most of the water vapor entering the Sonoran Desert at low-levels (below 700 mb) arrives from over the northern and central Gulf of California, with a slightly greater flux into the region occurring during the *dry* phase. Above 700 mb, moisture transported into the Sonoran Desert during both regimes is a mixture of water vapor from over the Gulf of Mexico and Gulf of California, and from residual convective inputs over the Sierra Madre Occidental mountains of Mexico. During *wet* periods, however, a longer fetch through the moist air mass above western Mexico results in a greater moisture flux into the Sonoran Desert aloft. Less water vapor from

over the Gulf of Mexico flows into western Mexico and the Sonoran Desert under wet conditions than during dry phases, both above and below 700 mb.

2 Introduction

Throughout the summer in the Sonoran Desert, the atmosphere typically undergoes several oscillations between hot, dry conditions with little rainfall, and slightly cooler, more humid weather with frequent afternoon thunder-showers. (A map of geographic regions referred to in this paper is included in Figure 1.) The transformation between *dry* and *wet* regimes can be gradual, spanning several days, or very abrupt, taking place within a 24 hour period (Bryson and Lowry 1955; Reitan 1957; Adang and Gall 1989; Watson *et al.* 1994). Such transitions in precipitation/moisture are illustrated in Figure 2, which shows the daily fraction of surface observing stations in southeast Arizona (State Climatological Division 07) reporting measurable precipitation during July and August, 1985 to 1992. (Arizona Climatological Divisions and locations of observing stations within Division 07 are identified in Figure 3). Both *intraseasonal* and *interannual* variability are evident. Despite its yearly and daily inconsistency, the summer monsoon produces 30-60% of the annual rainfall across the Sonoran Desert (Douglas *et al.* 1993).

In view of the large percentage of annual rainfall received during the summer monsoon, we believe that a thorough understanding of the mechanisms associated with *intraseasonal* variability may be a necessary prerequisite to achieving skillful seasonal precipitation forecasts. Unfortunately, studies dealing with variability of the North American Summer Monsoon are limited and mainly focus on Arizona, which lies well to the north of the heart of monsoon. Intraseasonal variability in the monsoon was first illustrated by Reed (1933), who showed a relationship between the northward movement

of the upper-level subtropical anticyclone and precipitation over the Desert Southwest. Variability of summertime rainfall was further documented by Bryson and Lowry (1955), while Reitan (1957) showed that variations in precipitation over Arizona are strongly correlated to the precipitable water over Phoenix. Hales (1972, 1974) and Brenner (1972) attributed the variability of humidity and rainfall to low-level gulf surges, sporadic northward surges of moisture from the Gulf of California into Arizona. Later diagnostic studies stressed the importance of the latitudinal position of the subtropical ridge in regulating mid-tropospheric moisture and precipitation. Synoptic climatologies for *wet* and *dry* monsoon periods, termed *bursts* and *breaks* by Carleton (1986), indicate that a ridge position north of the seasonal mean results in increased cloud cover and rainfall for Arizona; the opposite is true for southerly displacements (Carleton 1986; Carleton and Carpenter 1990). Watson *et al.* (1994) report similar results for *wet* and *dry* composites based on daily cloud-to-ground lightning totals in Arizona.

This study will focus on *intraseasonal* variability over the greater Sonoran Desert region of southwestern North America which, as we shall later see, overlaps the region of greatest variability in summertime convection. To address this issue, each day in July and August, 1985-1992, is classified as either *wet*, *dry*, or *transition*, based on the percentage of observing stations reporting precipitation in SE Arizona. Atmospheric conditions over Southwest North America and adjacent ocean environs are then compared for composite *wet* and *dry* monsoon regimes. We consider the variability of winds, specific humidity, precipitable water, convective instability, frequency

of cold cloud, and parcel trajectories. Although previous works have examined the variability of winds and specific humidity, to our knowledge no works exist that considered *intraseasonal* variations of the water vapor flux, parcel trajectories, and cloud top temperatures over this portion of the Desert Southwest.

As a continuation of the climatological study of Schmitz and Mullen (1995, hereafter referred to as SM), this work will employ the global analyses from the European Center for Medium-Range Weather Forecasting (ECMWF). The use of the ECMWF analyses provides several advantages over the data employed in prior research. Compared to *burst* and *break* synoptic climatologies based on other data (Carleton 1986; Watson *et al.* 1994), the ECMWF analyses possess higher spatial and temporal resolutions. Also, they provide vertical velocities consistent with model formulations and data assimilation procedures. Previous studies did not analyze this aspect of the flow field.

3 Data

3.1 ECMWF Analyses

Uninitialized global analyses produced by the ECMWF are used in this study. The analyses were obtained from the National Center for Atmospheric Research (NCAR). Analyses are available four times daily (0000, 0600, 1200, and 1800 UTC) for the surface and 14 mandatory levels from 1000 mb to 10 mb. The spectral truncation for the ECMWF analyses is triangular 106 (T106). The corresponding transform grid is approximately 1.125° latitude by 1.125° longitude.

For this study a subset of the global Gaussian grid, bounded approximately (to the nearest five degrees) by 5°N to 40°N and 90°W to 120°W , inclusive, is employed. Only analyses for the surface and eight lowest mandatory levels (1000, 850, 700, 500, 400, 300, 250, 200 mb) are used in this study. For a more thorough discussion of the ECMWF analyses as they relate to this study, the reader is referred to SM and references contained within.

3.2 Radiosonde Data

The radiosonde (RAOB) data employed in this study were extracted from NCAR Dataset DS390.1, United States Controlled TD56 Time Series RAOBS, which are prepared and maintained by the Data Support Section, Scientific Computing Division at NCAR. RAOB data are available for mandatory and significant levels up to 10mb. Each RAOB is hydrostatically checked. Twice daily (0000 and 1200 UTC) RAOBs for Tucson, Arizona for the period spanning July and August of each year, 1985-1992 were obtained. Composite soundings were constructed by first interpolating individual ones to every 10 mb (SFC, 910 mb, 900 mb, ...), and then averaging the isobaric data.

3.3 ISCCP Satellite Data

The infrared satellite imagery employed in this study are from the International Satellite Cloud Climatology Project (ISCCP) Archive. They are part of the ISCCP global, reduced-resolution, infrared and visible radiance (B3) data set possessing a horizontal resolution of $\sim 30\text{km}$. Infrared images are available every three hours for the years 1985 to 1991. This study uses 6

hourly infrared images for the synoptic hours of 0000, 0600, 1200, and 1800 UTC for July and August 1988-1990. The other summers were not available locally at the time of this study. To minimize data processing and storage requirements, only infrared pixels within a small subregion of the larger North American Satellite perspective, coincident with the regional ECMWF domain, are analyzed. For a more complete description of the ISCCP data products the reader is referred to Schiffer and Rossow (1985) and Rossow and Schiffer (1991).

4 Analysis Procedures

4.1 Selection Criteria for Wet and Dry Monsoon Periods

Douglas *et al.* (1993) convincingly show that Arizona and Sonora, Mexico are on the northern fringe of the primary monsoon region, that being the western slopes of the Sierra Madre Occidental where afternoon summertime precipitation is more consistent and extended *dry* periods during the rainy season are less evident. Only as one moves northward, into northwestern Mexico and southeastern Arizona, do monsoon *bursts* and *breaks* become more distinct. However, move much farther north, into central and northern Arizona, and the variability of precipitation once again decreases. This interpretation is supported by Figure 4, which shows the standard deviation of 0000 UTC cloud top temperatures (CTT) for July and August 1988-1990. The center of the greatest variability in late afternoon CTT occurs over the Sonoran Desert, to the northwest of the more persistent convective activity

over western Mexico.

In view of the fact that SE Arizona lies on the northern fringe of enhanced variability and that reliable precipitation records are more readily available for Arizona than for Mexico, daily precipitation data for Division 07 (i.e. southeast Arizona) of the Arizona Climatological subdivisions are employed (see Fig. 3) to identify *bursts* and *breaks* in the monsoon. *Wet* and *dry* days during July and August are defined as those days on which 50% or more, or fewer than 25% of the stations report measurable rainfall, respectively. This selection criterion categorizes 172 days as *wet*, 171 as *dry*, and 153 as *transition*. The average *wet* period lasted ~ 3 days, the average *dry* period ~ 4 days, and the average *transition* period ~ 2 days. Maximum lengths of *bursts*, *breaks*, and *transitions* were 10, 13, and 7 days, respectively. Results to be presented are composites for all *wet* and *dry* days.

4.2 Vertically-Integrated, Atmospheric Flux of Water Vapor

To illustrate the transport of atmospheric moisture under each monsoon regime, vertically-integrated water vapor flux vectors ($\vec{Q}$) are computed using

$$\vec{Q} = \langle \overline{q\vec{V}} \rangle = \int_{p_T}^{p_S} \left(\overline{q\vec{V}} \right) \frac{dp}{g} \quad (1)$$

where $\vec{V}$ is the horizontal velocity, q is the specific humidity, p_S is the surface pressure, and p_T is the pressure at the top of the atmosphere, or 200 mb in this case. The overbar indicates an average over all *wet* or all *dry* days. Vertical integration is denoted by brackets. All other quantities have their

usual meteorological meaning. Following the procedures detailed in SM, the horizontal moisture flux for the atmospheric layers above and below 700 mb, and the vertical flux connecting the two are computed for *wet* and *dry* regimes. Average moisture flux across regional boundaries for each monsoon phase are also computed according to the methods described in SM, only for different subregions. Vertical integrals are calculated by the trapezoid rule. All atmospheric fields are assumed to vary linearly with pressure between mandatory levels.

4.3 Lagrangian Perspective

Because parcel trajectories provide additional insight into the transport and origins of water vapor, backward and forward trajectories are computed for each monsoon phase. Trajectories are determined using the method of Reap (1972) by the numerical integration of

$$S(x, y, p) = \int_{t_0}^{t_1} \bar{V}_3(x, y, p, t) dt, \quad (2)$$

where t is time, t_0 is the starting time, S is the position of the parcel at the time t_1 , and $\bar{V}_3$ is the mean 3D velocity vector. *Dry* and *Wet* period trajectories are estimated using the composite average horizontal and vertical velocity fields at 0000, 0600, 1200, and 1800 UTC for the respective regime. Trajectories are computed using a 2 hour time step. Trilinear interpolation is used to determine velocities at positions between grid points/mandatory levels.

5 Results

5.1 Velocity and Moisture Fields

Figure 5 shows the time-mean winds at the surface for both monsoon phases. Overall, the two fields are quite similar, especially east of the Continental Divide and south of 20° N. The most notable difference is the weakening of the northwesterly flow west of Baja California and the westerly flow across Baja California and the Gulf of California as the monsoon moves from *dry* to *wet* phase.

More dramatic and larger scale changes are evident at 500 mb (Fig. 6). Under *dry* conditions the subtropical ridge axis lies south of its mean position (cf. SM, Fig A.3), almost directly above central Sonora. When positioned in this manner, strong southwesterlies characterize the flow over all of Arizona and extreme northwest Sonora, while easterly winds exist over all of Mexico and the adjoining eastern Pacific south of $\sim 28^\circ$ N. During *wet* conditions, however, the ridge axis is positioned north of the Sonoran Desert, and southeasterly winds occur over southern Arizona, western Mexico north of $\sim 20^\circ$ N, and the Gulf of California. Similar changes are found at all levels between 700 mb and 200 mb (not shown).

The vertical motion fields (Fig. 7) of *wet* and *dry* regimes are very similar. Although little change is seen throughout most of the domain, a finger of enhanced upward motion ($\omega < -0.5\mu\text{b s}^{-1}$) extends from Sonora northward into Arizona and New Mexico during *wet* periods. Subsidence occurs over extreme NW Mexico, SW Arizona, and the northern Gulf of California during

both regimes, but it is stronger and protrudes further eastward into Sonora and Arizona during *dry* periods.

5.2 Moisture Fields

A noticeable redistribution of atmospheric water vapor accompanies the changes in the winds. At the surface (Fig. 5), specific humidities (q) over the Sonoran Desert increase by more than $\sim 2.0 \text{ g kg}^{-1}$ during monsoon *bursts*. In fact, such increases cover all of Arizona, western New Mexico, southern Utah and southwestern Colorado. Much smaller, insignificant differences are found elsewhere.

A similar moistening occurs at 500 mb (Fig. 6). Under both regimes, the 500 mb moisture field is characterized by an elongated maximum that extends northward from the eastern Pacific Intertropical Convergence Zone (ITCZ) along the Sierra Madre Occidental into Arizona and New Mexico. Maximum moisture contents occur over the Sierra Madre Occidental and the Sierra Madre Del Sur during both regimes. The northern most extent of the wettest air ($q > 3 \text{ g kg}^{-1}$) lies just south of the subtropical ridge axis. Consequently, as the monsoon passes from *dry* to *wet* phase and the subtropical ridge moves into northern Arizona, the greatest moistening occurs over the Sonoran Desert. Mid-tropospheric specific humidities over this region are $\sim 1.0 \text{ g kg}^{-1}$ larger under *wet* conditions than during *dry* periods.

The precipitable water ($PW = \langle q \rangle$) fields (Fig. 8) indicate that the greatest values are found over the tropical East Pacific and SW Gulf of Mexico regions during both regimes. The *wet* – *dry* PW difference field (Fig. 9.a),

however, reveals that the largest changes are found over the SW United States and NW Mexico, with a maximum increase of ~ 1.2 cm (~ 0.5 inches) centered over the Sonoran Desert during the *wet* phase. No significant change is found east of $\sim 103^\circ\text{W}$ and south of $\sim 25^\circ\text{N}$.

If the changes in precipitable water for the layers above and below 700mb are considered separately (Fig's. 9b and c), we find that the moisture increase is distributed rather evenly between the two layers. Because of terrain elevation effects, the low-level increases are somewhat more localized than those aloft, with values above 0.5 g cm^{-1} being confined between the Mogollon Rim, the high terrain of northwest Mexico, and the northern Gulf of California.

Overall, the *wet* and *dry* period moisture and wind fields are in accord with prior results (Reed 1933; Reitan 1957; Carleton 1986; Carleton and Carpenter 1990; Watson *et al.* 1994b). At the surface, however, the flow remains predominantly westerly over the Gulf of California during both regimes, with only a slight enhancement of the southerly wind component over the gulf occurring during *bursts*. Thus the surface wind is almost 90° out of phase with the special observations of Badan-Dangon *et al.* (1991) and Douglas *et al.* (1993), whose results are based on much shorter periods which may not be representative of longer term conditions. Aside from that, the ECMWF analyses are felt to accurately represent the salient features of the *intraseasonal* variability for the region.

5.3 Convective Indices

As the atmosphere over the Sonoran Desert moistens, it also becomes more unstable. With average maximum surface temperatures varying little between monsoon regimes, the higher specific humidities during *bursts* produce an environment more conducive to convection. Assuming that conditions above Tucson are representative of the greater Sonoran Desert, *burst* and *break* composites of 0000 UTC radiosonde data (Fig. 10) indicate that dew point temperatures increase by $\sim 5^\circ \text{C}$ at all isobaric levels up to 300 mb during the *wet* phase. As a consequence, the lifting condensation level (LCL) for a surface parcel falls ~ 100 mb during this regime (Table 1). Most importantly, the convective available potential energy (CAPE) is positive during the *wet* regime, and it is much bigger than during the *dry* regime. Under *wet* conditions, surface parcels experience, on average, a 350 mb thick layer of modest positive buoyancy, extending from ~ 650 mb (the level of free convection LFC) to ~ 300 mb (the level of neutral buoyancy LNB). In addition, the moister atmosphere aloft further facilitates convective development by reducing the dissipating effects of entrainment. Despite the more favorable environment, surface heating alone probably cannot sustain widespread, deep convection; it seems likely that some lifting is also required. The -14°K difference in equivalent potential temperature Θ_e between the surface and 500 mb (the approximate level of minimum Θ_e) indicates the sounding is potentially unstable, but sufficient lifting of the entire air column is required to release this potential instability. Once precipitation commences, more buoyant energy can be released due to downdraft convective available potential

energy, or DCAPE (Emanuel 1994, p172). Although only a fraction of the DCAPE is usually realized (Emanuel 1994), it is interesting to note that the DCAPE for a parcel originating at cloud base (LCL) exceeds the CAPE for a surface parcel. In fact, the composite *wet* sounding resembles the *inverted-V* sounding of Beebe (1955) that is conducive to vigorous downdrafts (Bluestein 1993, p453-454).

The moister and more unstable atmosphere of the *wet* regime leads to a greater frequency of convection over the Sonoran Desert during these periods. This fact is illustrated by Figure 11 which shows the frequency at 0000 UTC of $CTT < -38^{\circ}C$ for each monsoon regime and significant differences between these fields. The $-38^{\circ}C$ threshold is commonly used as a surrogate for deep convection (Maddox *et al.* 1991; Douglas *et al.* 1993). During *dry* periods, frequent (> 0.25) convective activity is largely confined to the south $\sim 30^{\circ}$ N with a maximum over Sinaloa. When *wet* conditions exist, the frequency of convection increases by typically 10-35% over the Sonoran Desert. Little change in convective frequencies is observed over Sinaloa, however, the heart of the monsoon. These differences are consistent with the changes observed in the ECMWF vertical motion fields.

5.4 Vertically-Integrated, Atmospheric Flux of Water Vapor

If we compare the SFC-200 mb moisture flux for monsoon *bursts* and *breaks* (Fig. 12), statistically significant differences are found over the Sonoran Desert, Mexico, the Gulf of California, and the tropical East Pacific. During *dry* periods, the southward displaced subtropical ridge puts the Sonoran

Desert under the influence of a strong southwesterly flux which imports moisture from over the extratropical Pacific and northern Gulf of California into the region. South of $\sim 30^\circ$ N an easterly flux is found over Mexico and the Gulf of California, effectively confining tropical moisture south of this latitude.

Under *wet* conditions the northward displaced subtropical ridge results in a strong southeasterly flux over Mexico, north of 20° N, and the Gulf of California. Moisture over Mexico, the Gulf of California, and the tropical East Pacific appears to flow directly into the Sonoran Desert. Concurrently, the water vapor flux from over the extratropical Pacific and northern Gulf of California is weaker and restricted to the western portion of the Sonoran Desert.

To differentiate between the role of low-level and upper-level transports, water vapor fluxes for the SFC-700 mb and 700-200 mb layers are presented in Figures 13 and 14. Significant differences in the low-level flux are found over the greater Sonoran Desert region. During *dry* periods, a southwesterly flux blankets the entire Sonoran Desert, with most of the moisture coming from over the extratropical Pacific and northern Gulf of California. Much weaker, less organized fluxes are found over the central and southern Gulf of California, and coastal West Mexico. Low-level moisture from over the tropical East Pacific flows inland into Sinaloa but penetrates no further north than $\sim 25^\circ$ N.

When the monsoon is in its *wet* phase, fluxes over the northern Gulf of California shift to a more southerly direction, thereby covering a smaller

portion of the Sonoran Desert. Concurrently, less moisture from the extratropical Pacific flows across Baja California and into the northern Gulf region. Further south, a stronger, more organized southerly flux from the central Gulf of California into the southern Sonoran Desert is evident. A greater northward flux from the southern Gulf and tropical East Pacific into Sinaloa is also indicated.

The primary difference between 700-200 mb (Fig. 14) fluxes for *wet* and *dry* regimes involves the positioning of the subtropical ridge axis. During *breaks*, the ridge axis lies above the Sonoran Desert, resulting in a weak anti-cyclonic flow of moisture aloft. South of the ridge axis, a strong easterly flux extends across central and western Mexico, the southern Gulf of California, and into the subtropical Pacific. As a result, any 700-200 mb water vapor situated over Mexico or the southern Gulf region, south of $\sim 28^\circ$ N, does not flow over the Sonoran Desert during *breaks*.

During *bursts*, the subtropical ridge axis lies over central Arizona, along the northern limit of the Sonoran Desert. A broad area of upper-level southeasterly flux occurs over virtually all of Mexico north of $\sim 20^\circ$ N and the Gulf of California. This transport nearly parallels the Sierra Madre Occidental, with a slight deviation directed down the western slopes. As a result, upper-level moisture over Mexico appears to flood the Sonoran Desert from the southeast, while over the Gulf of California and tropical East Pacific it is directed away from the Sonora Desert.

Figure 14 indicates that upper-level moisture from above the Gulf of Mexico is transported into the Sonoran Desert regardless of the monsoon regime.

The greater southerly component in the transport field during *bursts*, however, indicates that less Gulf of Mexico moisture crosses the Mexican Plateau during *wet* periods, flowing northwestward along the high terrain instead of across it. During *dry* periods, the flux is more normal to the mountains, suggesting a greater westward transport across the Sierra Madre Occidental and Mexican Plateau.

5.5 700 mb Vertical Flux

Figure 15 illustrates the *wet* and *dry* period vertical moisture flux and the difference between these fields. Both monsoon regimes are characterized by an elongated band of upward moisture transport that extends from the ITCZ along western Mexico before reaching Arizona and New Mexico. It is surrounded on both sides by generally weaker descent.

Consistent with our selection criterion, enhanced upward moisture transport exists over SE Arizona during monsoon *bursts*. Comparable increases are found over Sonora. Thus, assuming the additional moisture is not rained out, more water vapor is available for transport aloft during *wet* periods. No significant change occurs in the vertical flux over Sinaloa, however.

5.6 Moisture Transport across Regional Boundaries

As a means for quantifying the variations in the water vapor transport between *wet* and *dry* phases, differences in the large scale moisture among four subregions are considered. These subregions are illustrated and defined in Figure 16. They were chosen to delimit the northern and southern portions

of the Sonoran Desert, the Gulf of California, and the region of persistent convective activity over western Mexico. Only the moisture flux above and below 700 mb and the vertical flux connecting the two will be considered.

Comparison of the *burst* and *break* regional fluxes (Fig. 17) confirms many of the points discussed above. During the *dry* regime moisture above 700 mb is transported westward through the entire domain south of $\sim 30^\circ$ N, while eastward transport occurs to the north of this latitude. Below 700 mb a westerly flux dominates the transport over Baja California Norte and extends through the Sonoran Desert. A southerly transport from the tropical East Pacific exists within both layers, but most of the moisture entering the southern GC subregion is transported westward by the prevailing easterly flow.

Once the *wet* regime is in place, however, several significant changes occur. Foremost is the reduction of the water vapor transport into the domain from the east, indicating that less Gulf of Mexico moisture crosses the continental divide during the *wet* phase, both above and below 700 mb. The decrease in water vapor transported into MXW from the east is compensated by other processes. Above 700 mb, the shift from *dry* period easterlies to *wet* period southeasterlies over Mexico and the Gulf of California not only decreases the amount of moisture entering MXW_U from the east, but it also reduces the amount transported westward through MXW_U by an almost equal amount. Below 700 mb, the sense of the transport across the MXW_L-GC_L border reverses and now flows into MXW_L. This moisture input, combined with the reduction in the westward transport from MXW_L into the tropical East

Pacific, more than counterbalances the decreased input from over the Gulf of Mexico.

Another important feature is the apparent constancy of the low-level moisture flux into the Sonoran Desert from over the GC_L . Whereas most of the low-level water vapor over NSD_L originates from over the northern GC_L under both monsoon regimes, the primary source for the SSD_L depends on the phase. During *breaks* it receives most of its moisture from above the northern GC_L , but during *bursts* most comes from above the central GC_L region. As a whole, slightly more water vapor flows into the Sonoran Desert from above the northern and central GC_L region during *dry* periods.

Unlike the horizontal water vapor transports, little change is found in the vertical flux through the 700 mb layer. The only significant differences in the *wet* and *dry* period fluxes occur over the Sonoran Desert and act to transport more moisture upward.

5.7 Lagrangian Perspective

To better identify the origins of moisture transported into the Sonoran Desert, 3D parcel trajectories were computed for the composite-mean conditions of each regime. Swarms of four day back and forward trajectories were analyzed. Of the multitude of trajectories examined, only a small subset that are representative of conditions for the region are presented.

Forward trajectories for parcels originating at the surface of the Gulf of California for *wet* and *dry* periods are shown in Figure 18. They indicate that water vapor over the Gulf is carried into western Mexico and the Sono-

ran Desert regardless of the monsoon regime, with only a slight northward deviation being evident for *wet* trajectories. As these parcels move inland, they rise mainly due to forced orographic ascent. During *wet* periods, the parcels starting over the central Gulf turn northward, toward the Sonoran Desert as they rise. Under *dry* conditions they do not turn northward.

Parcels over the central Gulf at 850 mb (Fig. 19) also generally ascend to the north during *wet* conditions, but descend southward during *dry* periods. Those at 850 mb over the northern Gulf of California, however, move north-eastward through the Sonoran Desert during both phases of the monsoon. For parcels originating both at the surface and 850 mb over the northern Gulf, the *dry* period parcel moves farther to the east than the *wet* period parcel.

Parcels originating at 850 mb over the western slopes of the Sierra Madre Occidental (Fig. 20) are transported towards the Sonoran Desert during the *wet* monsoon phase. When the *dry* regime is in place, parcels south of $\sim 30^\circ$ N head toward the southwest, while those north of this latitude move to the northeast.

Figure 21 shows back trajectories terminating at 500 mb over the northern Gulf of California region and southern Arizona. During *dry* regimes, parcels descend into the Sonoran Desert from the southwest, sinking as much as 150 mb in four days. When the subtropical ridge is positioned to the north and the monsoon is in its *wet* phase, parcels stream into the Sonoran Desert from over the Sierra Madre Occidental of northwest Mexico, ascending at a rate of ~ 50 mb day⁻¹ along the entire path.

6 Discussion and Conclusions

In this paper the *intraseasonal* variations associated with the summertime North American Monsoon were investigated. *Wet* and *dry* monsoon periods were defined from rain gauge data for southeast Arizona. Conditions during *wet* days were compared to conditions during *dry* days. Cloud top temperatures, wind, specific humidity, precipitable water, convective instability, moisture flux, and parcel trajectories were examined. Our primary findings are as follows:

- Intraseasonal variability in cloudiness and convection, as inferred from satellite imagery, is greatest over the Sonoran Desert. The region of enhanced variability lies to the northwest of the region of most persistent convection and greatest rainfall, i.e. the western slopes of the Sierra Madre Occidental.
- More water vapor exists above the Sonoran Desert under *wet* conditions than during *dry* spells, as precipitable water increases by as much as ~ 1.2 cm (~ 0.5 inches). The additional moisture is fairly evenly divided between the SFC-700 mb and 700-200 mb layers over the Sonoran Desert, but a more widespread increase occurs above 700 mb.
- Deep convection, as judged from infrared satellite imagery, occurs with a much greater frequency over the Sonoran Desert during *wet* periods than during *dry*. During *breaks*, deep convection is generally confined south of $\sim 30^\circ$ N. During *bursts*, convection pushes northward into Ari-

zona and New Mexico. Concurrently, the upward moisture flux over Sonora and east-central Arizona increases significantly.

- During *dry* periods the subtropical ridge is displaced southward of its climatological position. A mixture of upper-level moisture from over the Gulf of Mexico, northern Gulf of California, and residual convective inputs over Mexico circulates anticyclonically throughout the Sonoran Desert. Strong subsidence over the Gulf of California and the southwestern Sonoran Desert is present. During *dry* periods, little or no upper-level moisture south of $\sim 28^\circ$ N over Mexico and adjoining Pacific reaches the Sonoran Desert.
- During *wet* periods, the ridge lies north of its mean position and upper-level moisture ascends into the Sonoran Desert from the southeast, bringing with it water vapor from the atmosphere above western Mexico. As during the *dry* phase, this moisture is a mixture of water vapor from over the Gulf of Mexico, and residual convective inputs over the Sierra Madre Occidental. However, the *wet* period upper-level flow has a longer fetch through the moist air mass over the Sierra Madre Occidental than it does during *dry* periods. Consequently, more residual moisture over western Mexico enters the Sonoran Desert aloft during *wet* periods. Upper-level moisture over the Gulf of California and tropical East Pacific, on the other hand, is steered to the west of the Sonoran Desert by the *wet* period upper-level winds.

- During *dry* periods, low-level moisture from above the southern Gulf of California and tropical East Pacific cannot reach the Sonoran Desert at any level. During *wet* phases, however, this low-level tropical moisture flows northward and inland along the east coast of the Gulf of California.
- The low-level transport of water vapor into the Sonoran Desert from above the northern and central Gulf of California is largely independent of monsoon phase. During both regimes the low-level flux is oriented from this region into southern Arizona and northern Sonora. In fact, slightly more moisture from over the central and northern Gulf of California enters the Sonoran Desert during the *dry* phase.
- Gulf of Mexico moisture is transported into the Sonoran Desert under both monsoon regimes, but less crosses the Mexican Plateau and enters the Sonoran Desert during *wet* conditions.

Since the amount of low-level water vapor flowing into the Sonoran Desert from over the Gulf of California remains virtually constant during the *wet* and *dry* regimes, factors other than the low-level flux must regulate widespread convection over the region. Our results suggest that precipitation depends critically upon the amount of upper-level moisture and the vertical motion. The northward movement of the subtropical ridge and concomitant southeasterly midtropospheric flux during *wet* conditions, all of which are related to changes in the large-scale flow pattern, prompts moisture aloft to move northward from the region of persistent convection over western Mexico. The

moistening of the middle and upper troposphere over the Sonoran Desert both makes the atmosphere more unstable and reduces the dissipating effects of entrainment. Concurrently, weak ascent occurs within this potentially unstable atmosphere. This mix of ingredients (persistent, low-level moisture flux from the Gulf of California; moistening of the mid-troposphere due to southeasterly flux; convectively unstable atmosphere; ascent) results in a greater frequency of convection during the *wet* regime.

Although our analysis suggests that the Gulf of California is an important source of low-level moisture for the Sonoran Desert, its true impact is difficult to infer from the T106 ECMWF analyses alone. As Schmitz and Mullen (1995) discuss, the Gulf of California and the terrain of the Baja Peninsula are not adequately resolved at a T106 spectral truncation. Finer resolutions are required to properly resolve the local geography. We believe that more mesoscale modeling efforts (e.g. Stensrud *et al.* 1994) and/or special field programs (e.g. Meitin 1991) are needed to determine the role of the Gulf of California.

The methodology employed in this study leads to a single, “average” pattern associated with *wet* conditions. Yet it is important to recognize that there can be noteworthy deviations about the *wet* composite which also yield widespread rainfall over the Sonoran Desert. In fact, it is quite possible that any individual day with widespread rain over the Sonoran Desert can vary greatly from the *wet* composite (e.g. Watson *et al.* 1994). McCollum (1993), using a much smaller sample size than ours (31 verses 172), subjectively identifies three basic patterns associated with severe thunderstorms over central

Arizona. One of his patterns (McCollum 1993, p.28) closely resembles the *wet* pattern, but his other two patterns deviate notably from it. The fact that there is some agreement between two studies indicates that the *wet* pattern is robust and synoptically meaningful. By employing statistical techniques such as cluster analysis, those days/patterns that deviate substantially from the *wet* composite can be isolated and further stratified into more meaningful synoptic patterns.

It is clear that significant differences exist between the *wet* and *dry* regimes of the North American Monsoon. However, the present study offers little insight into mechanisms associated with transitions between them. As previously noted, the subject has received only limited attention, and those studies that exist typically consider case studies (e.g. Reed 1933, Bryson and Lowry 1955, Adang and Gall 1989). Future diagnosis of multi-year data sets and of extended regional model simulations may provide insight into regime transitions associated with monsoon.

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Table 1: Convective Indices

	Wet	Dry
LCL	717 mb	614 mb
LFC	651 mb	543 mb
LNB	284 mb	321 mb
CAPE	346 J kg ⁻¹	46 J kg ⁻¹
Lifted Index	-2 K	0 K
$\Theta_e(500 \text{ mb-SFC})$	-14 K	-7 K

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Figure Captions

Figure 1. Geography of the SW North American region showing locations mentioned in the text. Stippled region indicates portion of the Sonoran Desert emphasized in this analysis. We have excluded that portion over the Baja peninsula. The greater Sonoran Desert, or Desert Southwest will be used to refer to the broader region including and surrounding the Sonoran Desert.

Figure 2. Daily Percentage of Surface Observing Stations in Southeast Arizona, State Climatological Division 07, Reporting Measurable Precipitation During July and August, 1985-1992. Division 07 contains 44 to 46 observing stations throughout the eight year period.

Figure 3. Arizona Climatological Divisions. Positions of observing stations in Division 07 are marked with a dot.

Figure 4. Standard Deviation of 0000 UTC Cloud Top Temperatures $< -38^{\circ}$ Celsius for July and August, 1988-1990.

Figure 5. Average winds and specific humidities at the surface for (a) dry and (b) wet periods. Contour interval for q is 2 g kg^{-1} ; regions of $q > 12 \text{ g kg}^{-1}$ are stippled. Maximum wind vector is 7.3 m s^{-1} in (a) and 6.4 m s^{-1} in (b).

Figure 6. Average winds and specific humidities at 500 mb for (a) dry and (b) wet periods. Contour interval for q is 0.25 g kg^{-1} ; regions of $q > 3 \text{ g kg}^{-1}$ are stippled. Maximum wind vector is 8.7 m s^{-1} in (a) and 8.2 m s^{-1} in (b).

Figure 7. Average vertical velocity for at 500 mb for (a) dry, and (b) wet periods. Contours are every ± 0.2 , ± 0.4 , ± 0.6 , ..., $\mu\text{b s}^{-1}$; values $< -0.2 \mu\text{b s}^{-1}$ are stippled.

Figure 8. Average precipitable water for the surface to 200 mb column for (a) dry, and (b) wet periods. Contour interval is 0.5 cm; values > 4 cm are stippled.

Figure 9. *Wet period – Dry period* difference in precipitable water for (a) the surface to 200 mb layer, (b) the 700 to 200 mb layer, (c) the surface to 700 mb layer. Contour interval is 0.10 cm. Only regions with differences that are significant at the 5% level are contoured. Statistical significance was determined using the standard two-tailed t-test for the differences between the means of two populations.

Figure 10. Composite soundings at Tucson, Arizona for (a) wet, and (b) dry periods. Soundings are plotted on Skew-T log-P Diagram. Each half wind barb equals 5 knots.

Figure 11. Frequency of Cloud Top Temperatures $< -38^{\circ}$ Celsius during July and August 1988-1990 for composite (a) dry periods, (b) wet periods, and (c) differences between wet and dry periods. Frequency is expressed as the fraction of *Dry* or *Wet* observations with CTT $< 38^{\circ}$ C, respectively. Contour interval is 5%. Only regions with differences that are significant at the 5% level are contoured.

Figure 12. Water vapor flux for the atmosphere extending from the surface to 200 mb for (a) Dry periods, (b) Wet periods, and (c) *Wet Period –*

Dry Period. Units are $10^2 \text{ g cm}^{-1} \text{ s}^{-1}$. Maximum vector is (a) $32.4 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$, (b) $30.4 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$, and (c) $8.8 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$. Only those differences significant at the 5% level are presented. Statistical significance of all vector differences was estimated using a two-tailed t-test on the differences between the *wet* and *dry* means, where the variance of each population was determined from the component of the instantaneous *wet* (*dry*) vector parallel to a unit vector in the direction of the mean *wet* (*dry*) vector.

Figure 13. Water vapor flux for the atmosphere extending from the surface to 700 mb for (a) Dry periods, (b) Wet periods, and (c) *Wet Period – Dry Period*. Units are $10^2 \text{ g cm}^{-1} \text{ s}^{-1}$. Maximum vector is (a) $25.9 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$, (b) $24.6 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$, and (c) $4.3 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$. Only those differences significant at the 5% level are presented.

Figure 14. Water vapor flux for the atmosphere extending from the 700 to 200 mb for (a) Dry periods, (b) Wet periods, and (c) *Wet Period – Dry Period*. Units are $10^2 \text{ g cm}^{-1} \text{ s}^{-1}$. Maximum vector is (a) $9.0 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$, (b) $8.7 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$, and (c) $3.6 \cdot 10^2 \text{ g cm}^{-1} \text{ s}^{-1}$. Only those differences significant at the 5% level are presented.

Figure 15. Vertical flux through the 700 mb level for (a) *Dry* periods, (b) *Wet* periods, and (c) *Wet Period – Dry Period*. Contour interval is $0.20 \text{ g cm}^{-2} \text{ day}^{-1}$. Only regions with differences that are significant at the 5% level are contoured.

Figure 16. Geographical areas used for the calculation of moisture transport across regional boundaries. Subregions are defined as the Northern

and Sonoran Desert (NSD), and Southern Sonoran Desert (SSD), respectively. Mexico West (MXW) of the Mexican highlands, and the Gulf of California (GC). Subregions above 700 mb are given the subscript U, for upper layer, and those below 700 mb the subscript L, for lower layer.

Figure 17. *Wet period – Dry period* regional boundary flux differences for NSD, SSD, MXW, and GC for the layers above and below 700 mb. All differences are presented as positive numbers. Arrows indicate the direction of the difference along each face of the volume. Numbers in parenthesis indicate the net Dry period flux across the corresponding boundary where a “–” sign indicates that the direction of the dry period flux is in the opposite sense as the difference vector. Underlined values refer to vertical transports. Magnitudes are presented as 10^{16} grams per month. Only those differences significant at the 5% level are presented.

Figure 18. Four day forward trajectories for parcels originating at the surface of the Gulf of California for (a) dry, and (b) wet, periods. Isobaric level of the parcel is indicated every 2 days. Arrow heads are plotted at the end of each one day back trajectory segment. Only trajectories started at 1200 UTC are shown. Trajectories were not found to be sensitive to the starting synoptic hour.

Figure 19. As in Fig. 18 except for parcels originating at the 850 mb level over the Gulf of California.

Figure 20. As in Fig. 18 except for parcels originating at the 850 mb level over western Mexico.

Figure 21. As in Fig. 18 except for four day back trajectories for parcels originating at 500 mb over the Sonoran Desert.

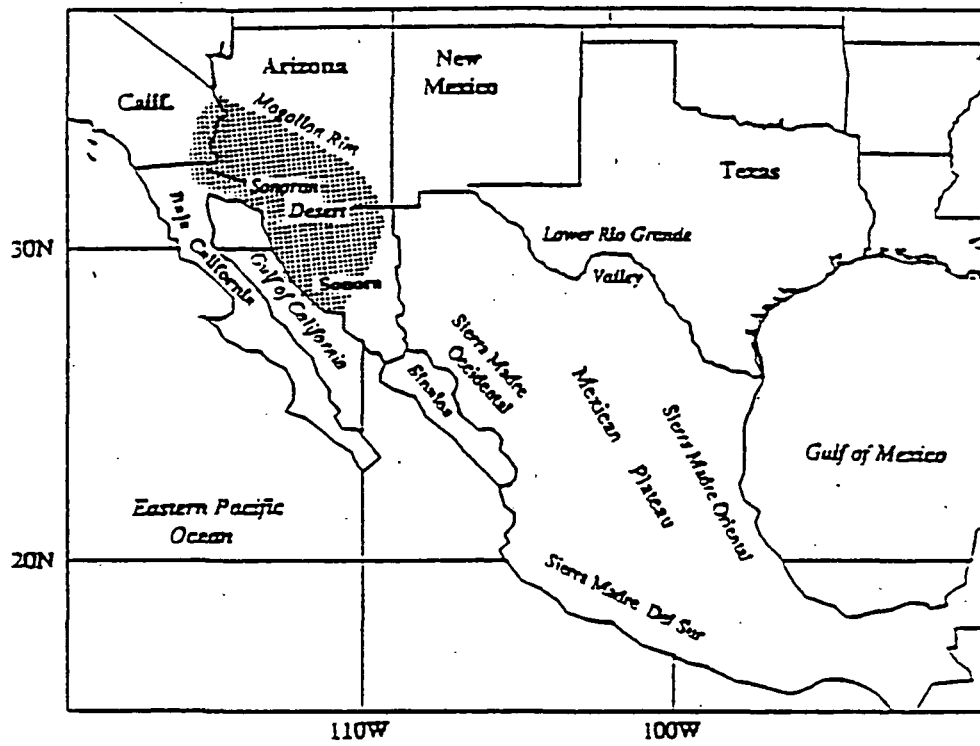


Fig. 1

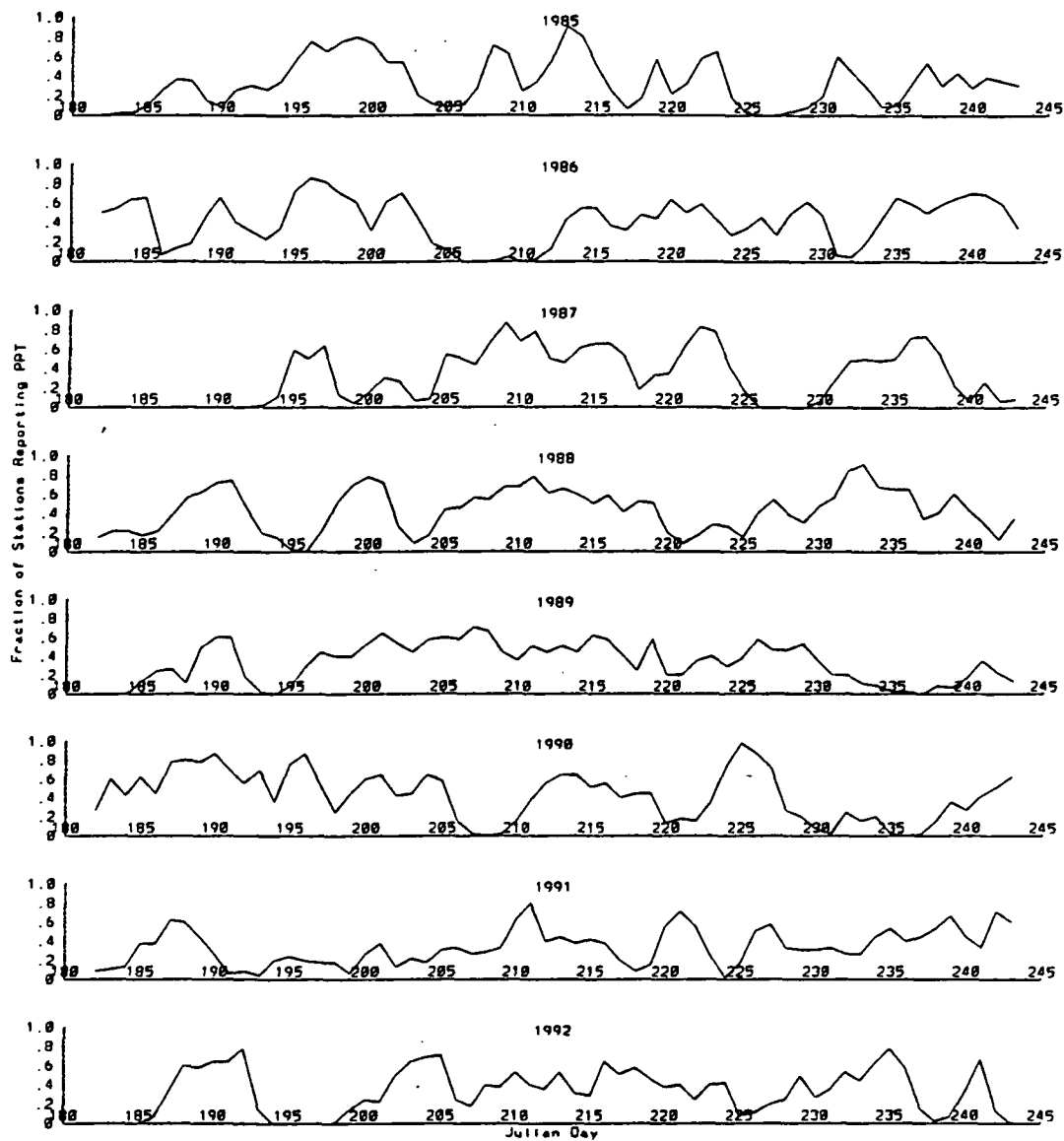


Fig 2.

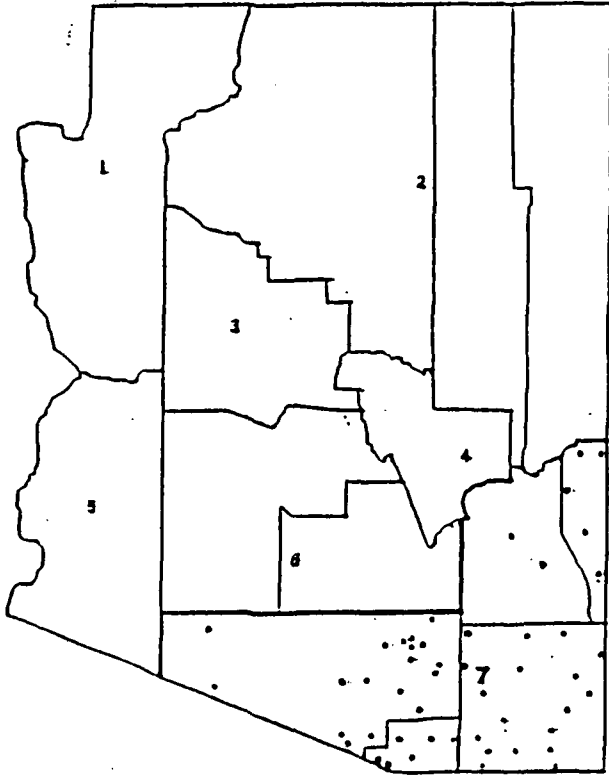


Fig 3

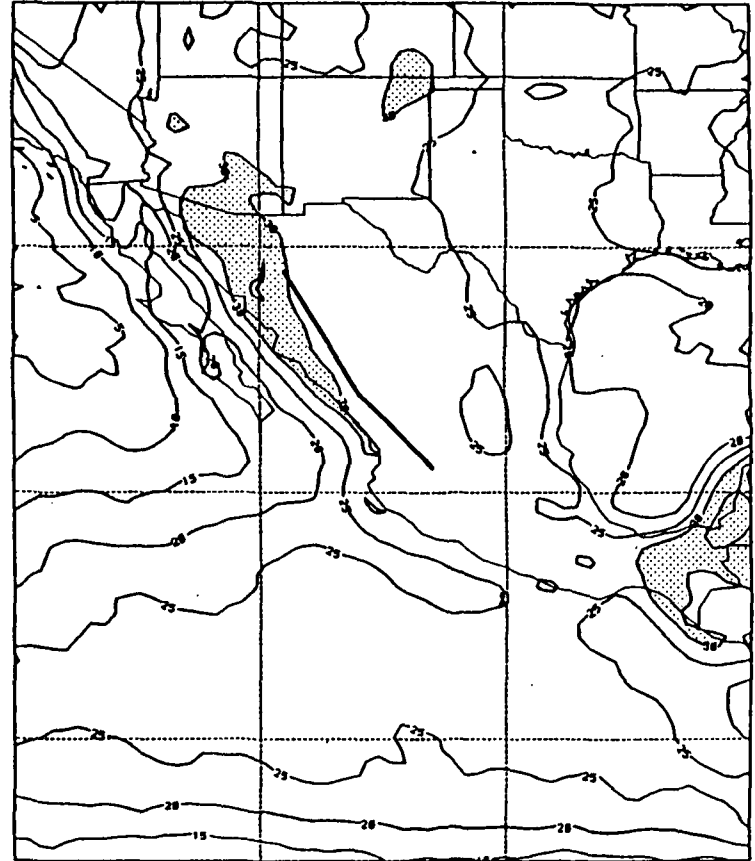


Fig 4.

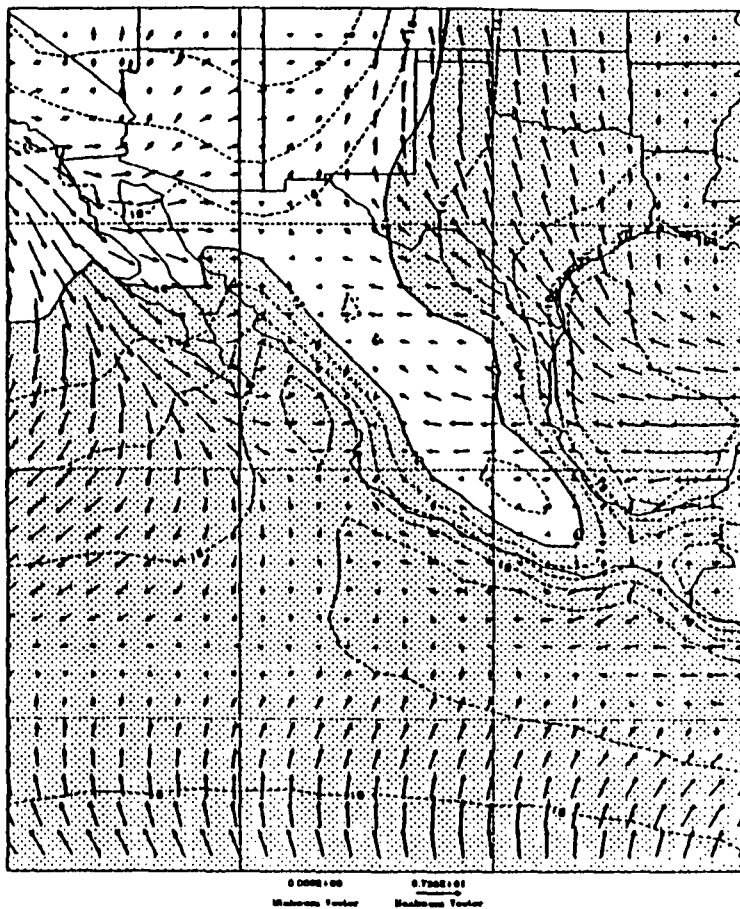


Fig 5a

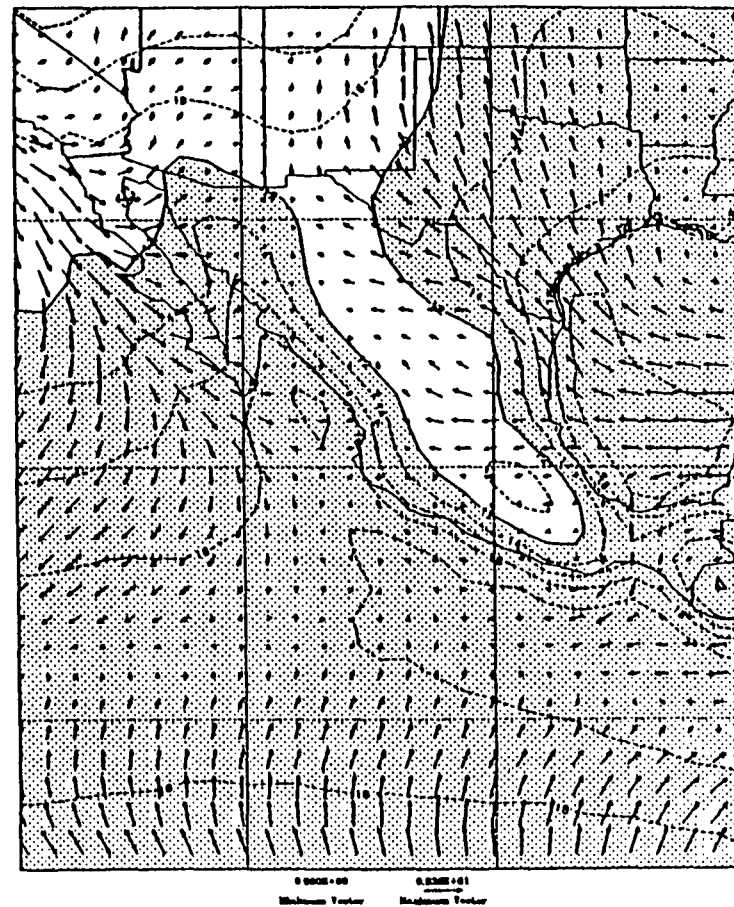


Fig 5.b

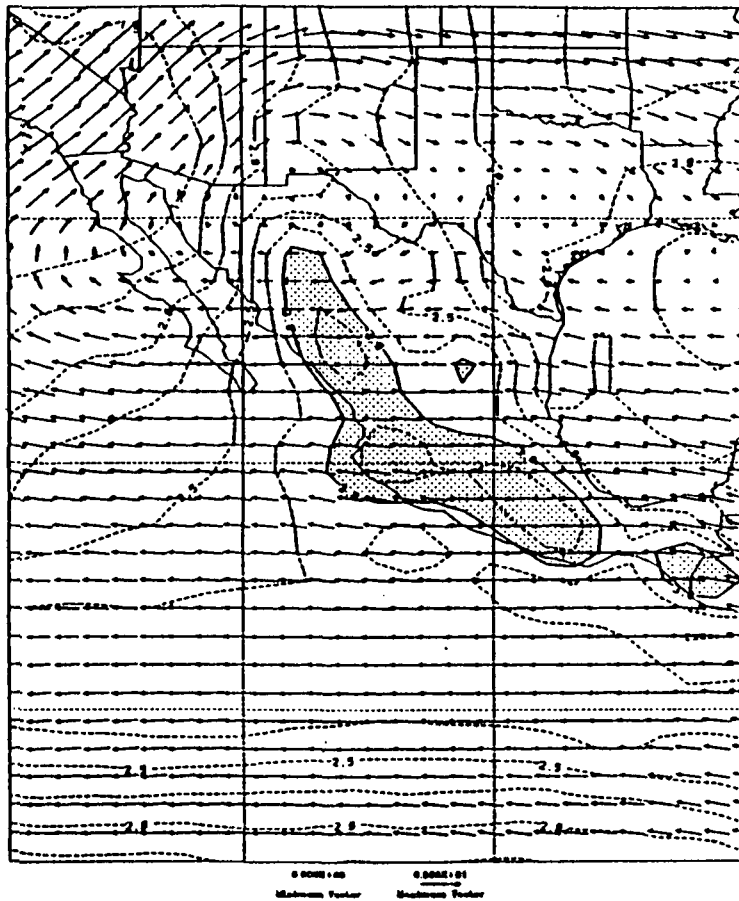


Fig 6.4

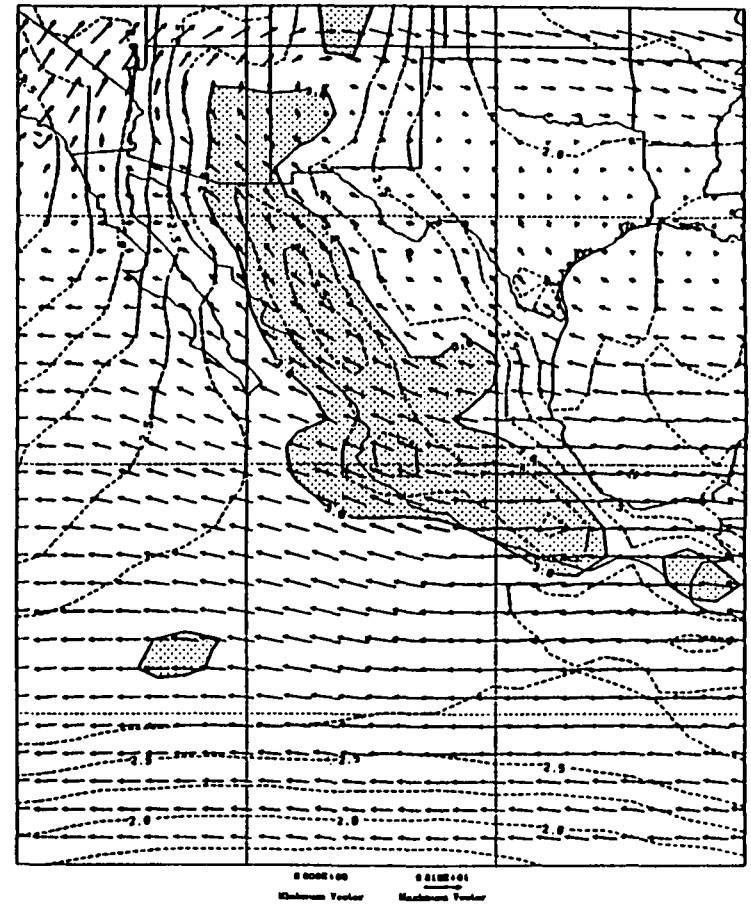


Fig 6.6

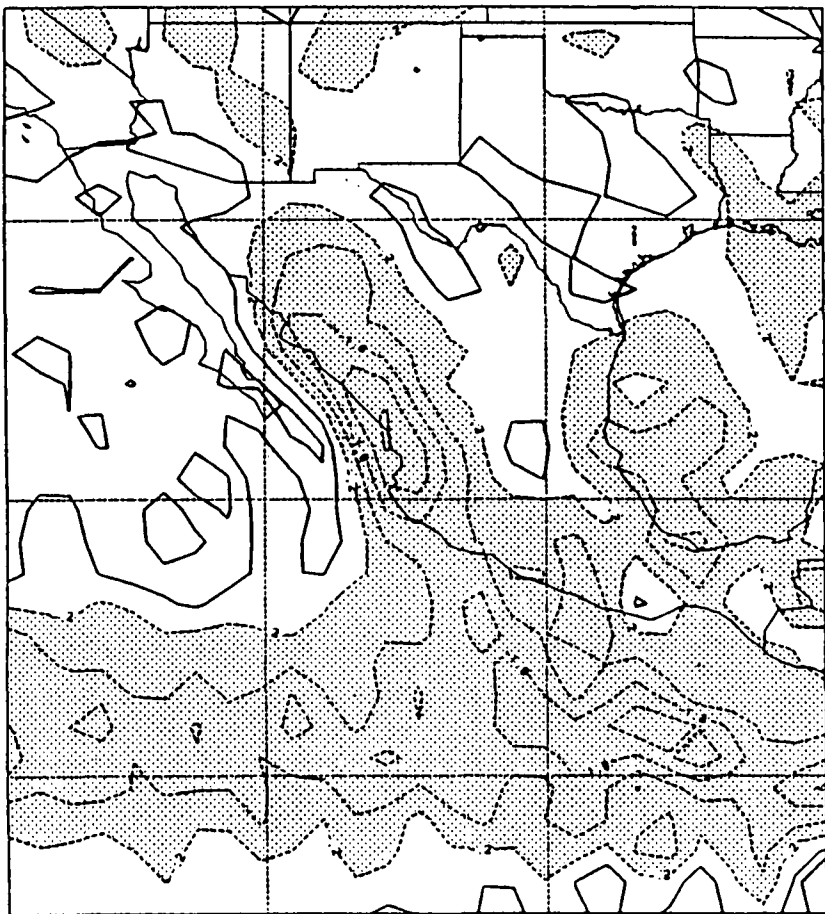


Fig 7.a

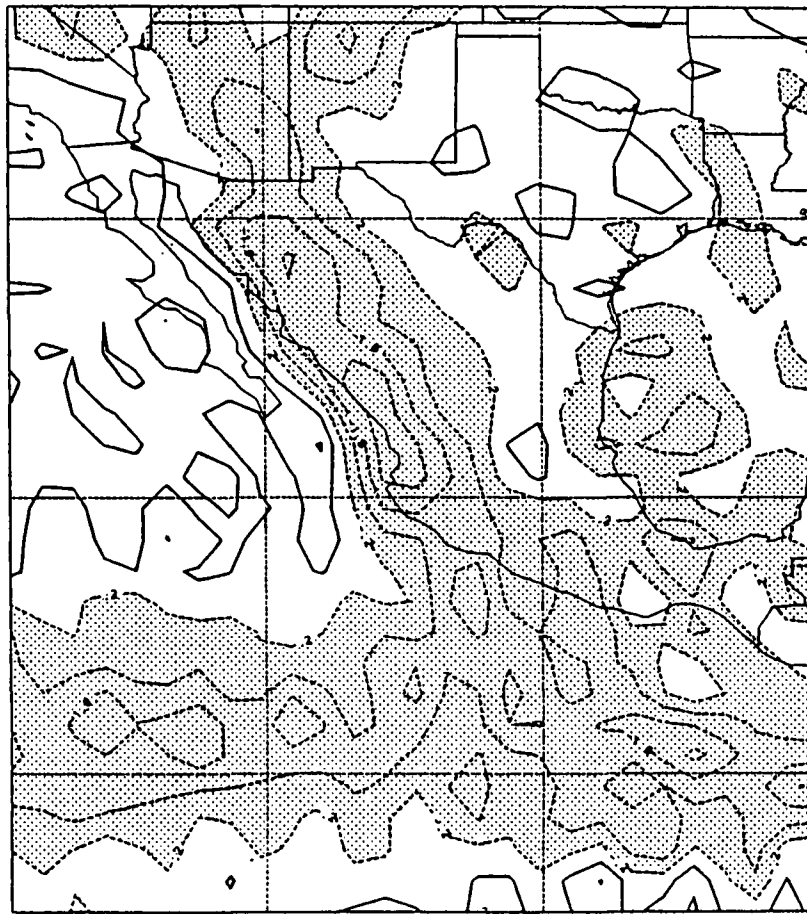


Fig 7.b

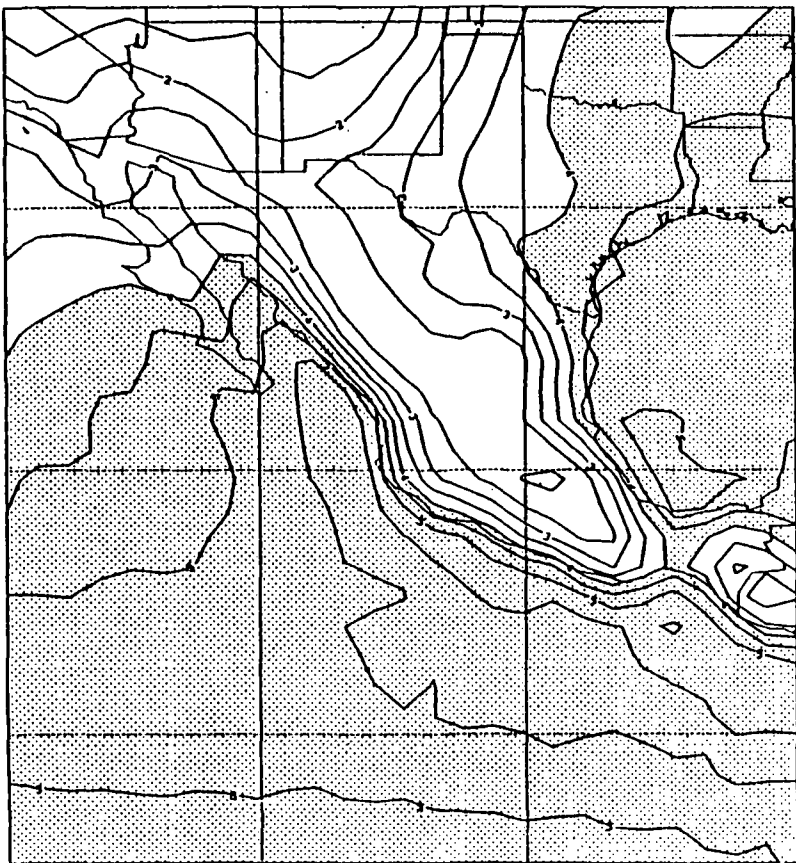


Fig 8.4

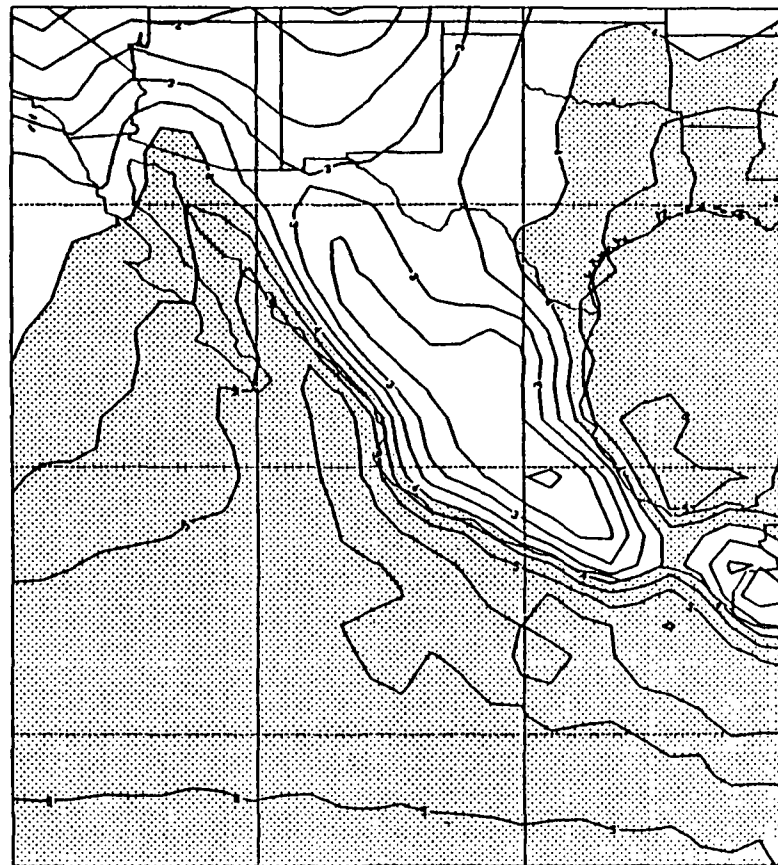


Fig 8.6

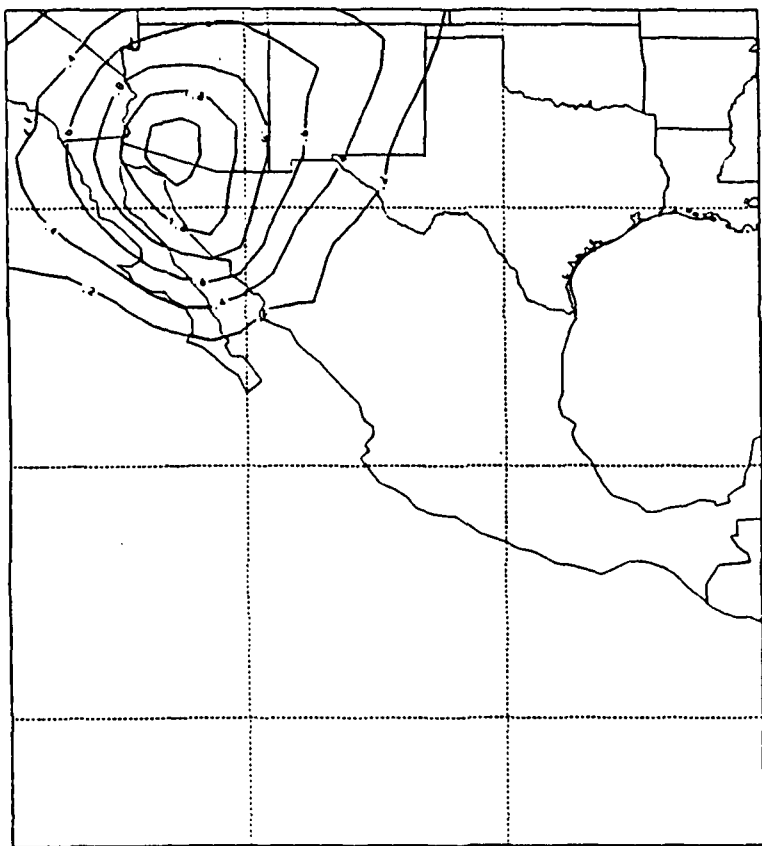


Fig 9.a

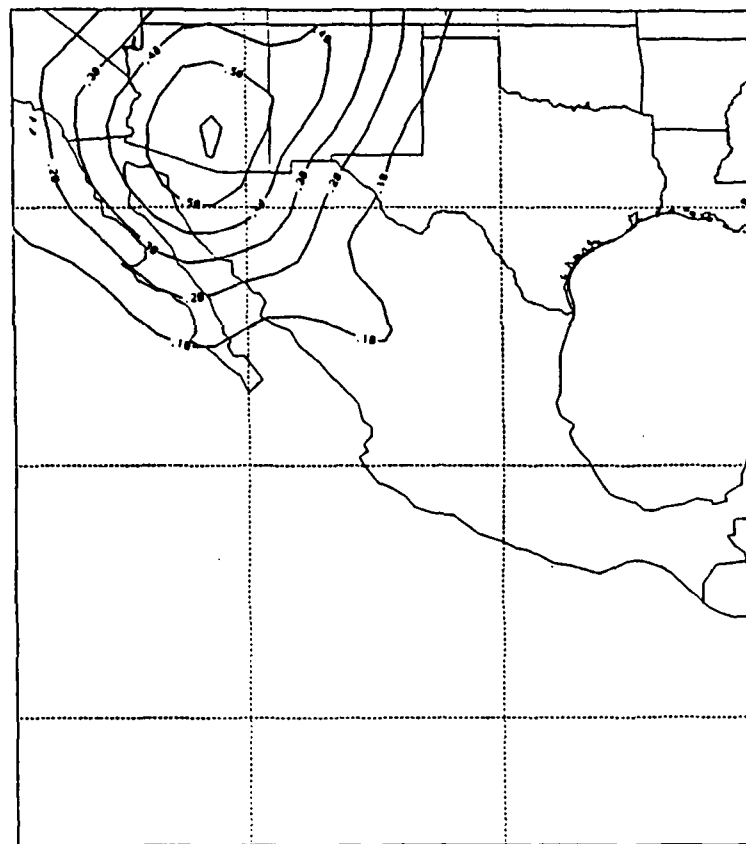


Fig 9.b

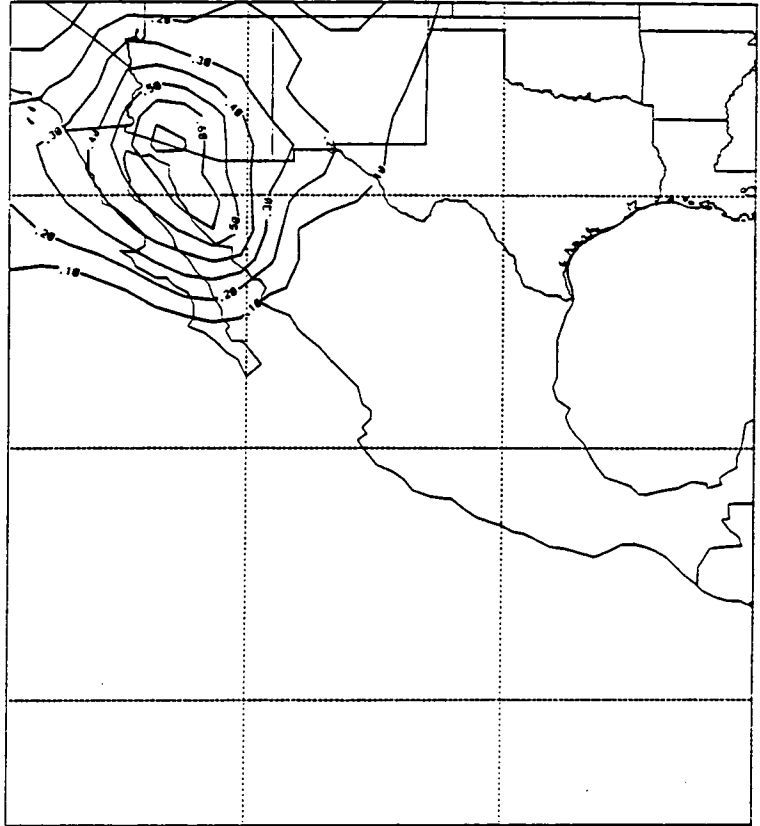
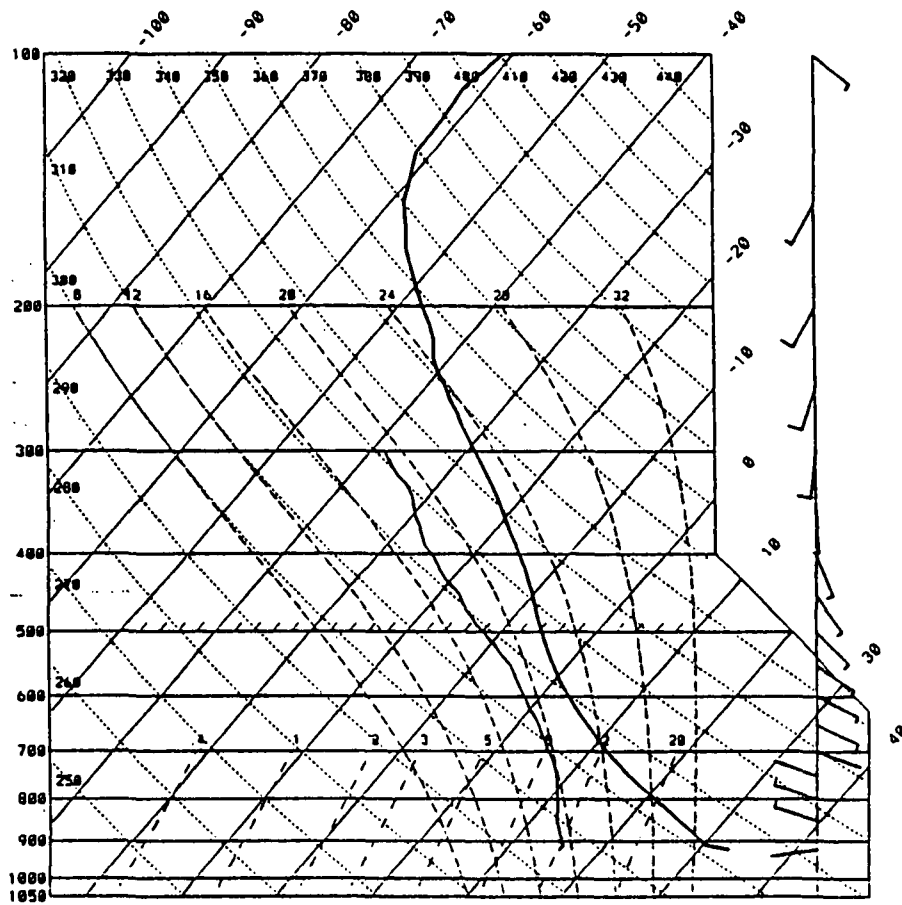
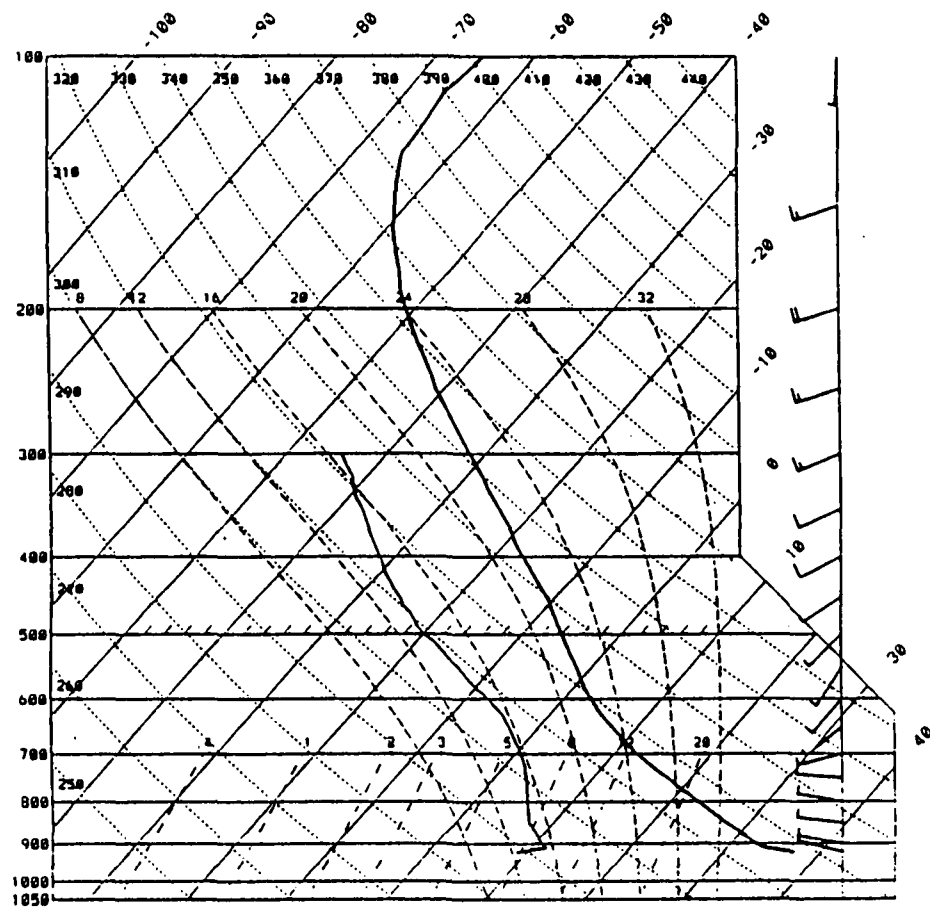
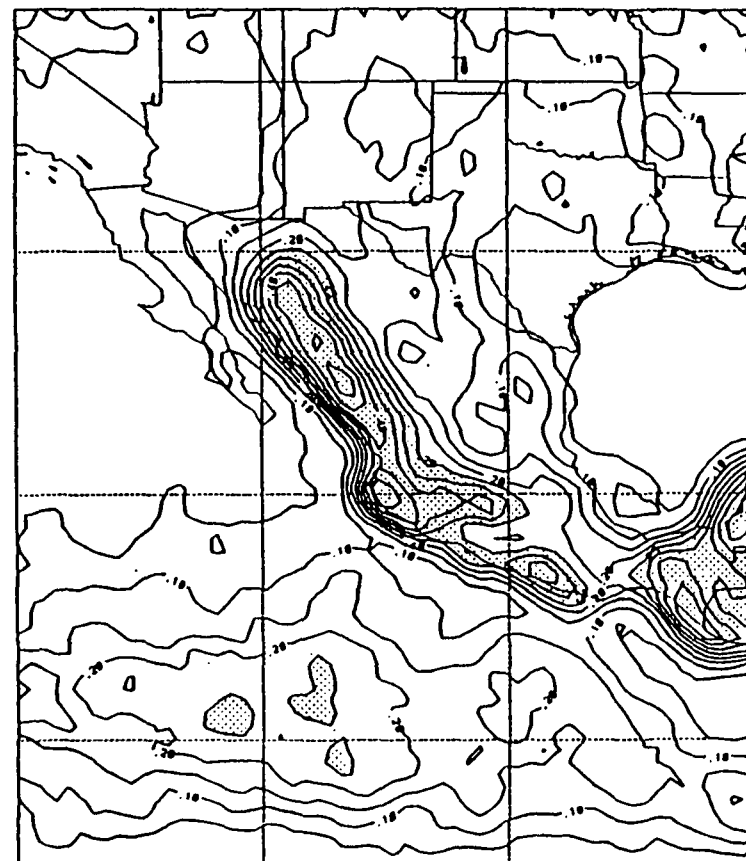
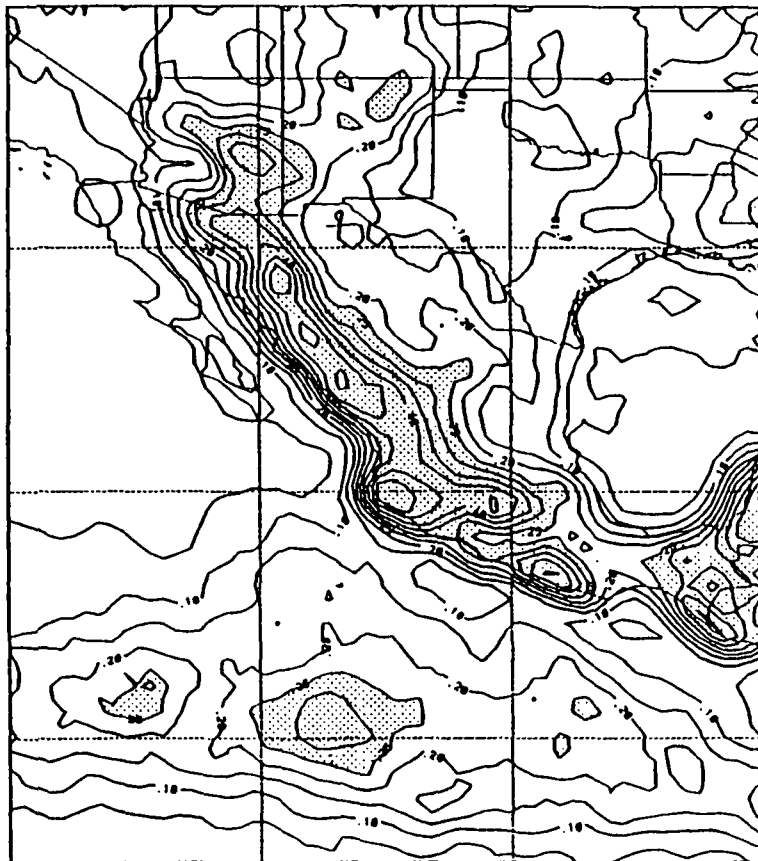


Fig 9.c





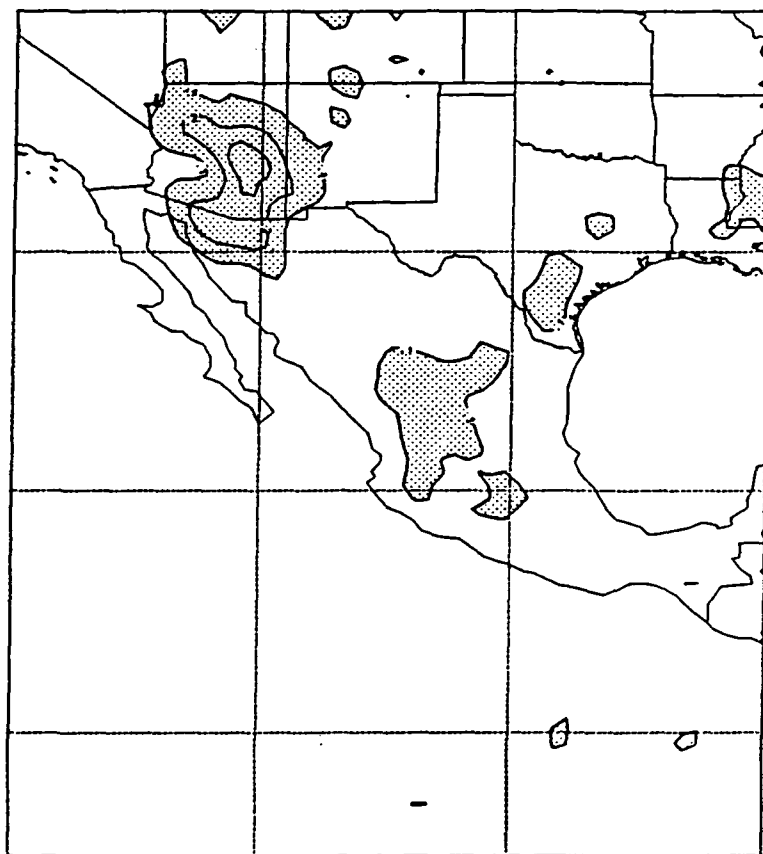
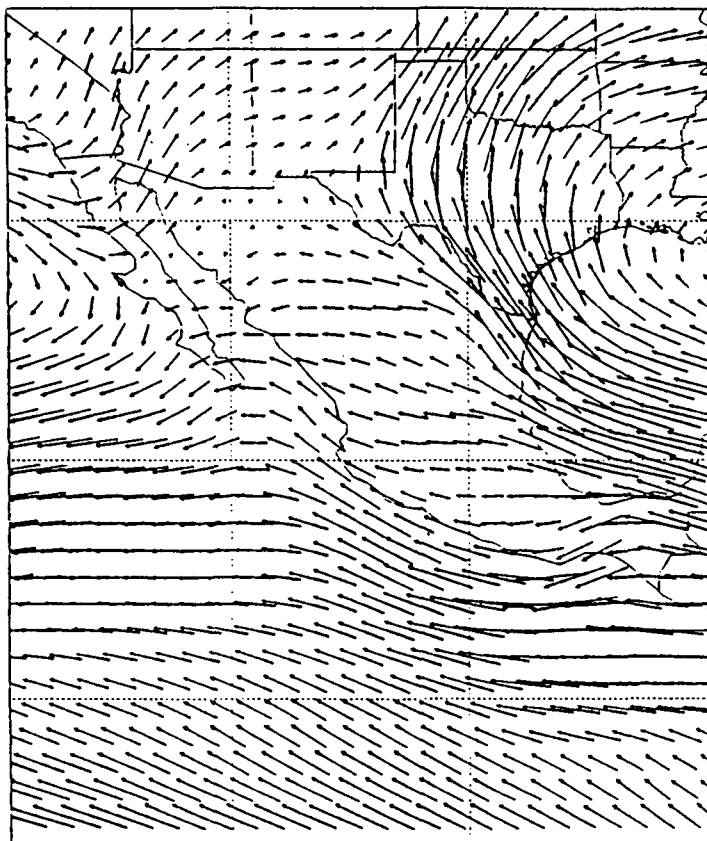
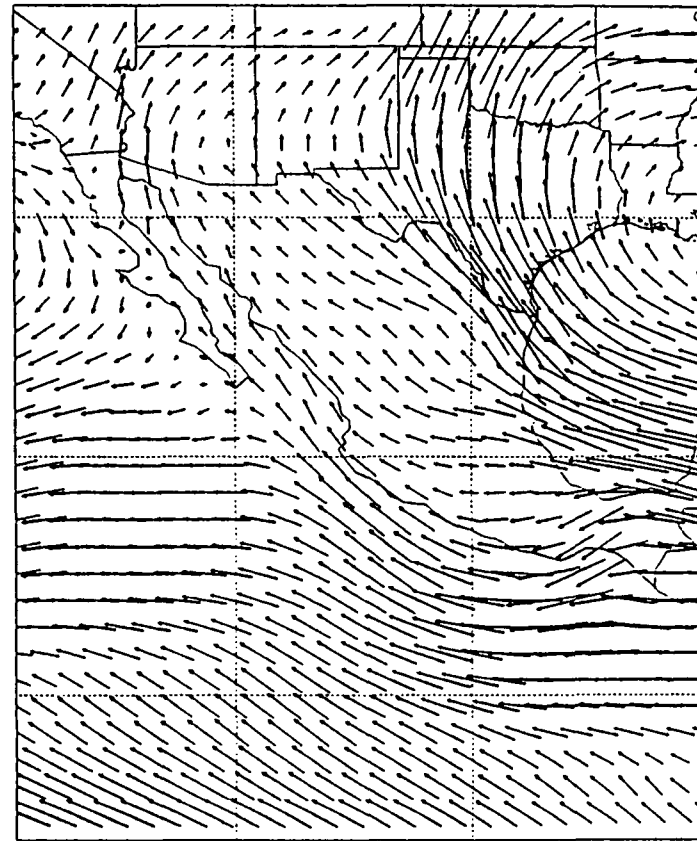


Fig 11.c



0.0000×10^0 1.2516×10^2
 Minimum Vector Maximum Vector

Fig 12.a



1.0000×10^0 0.3045×10^2
 Minimum Vector Maximum Vector

Fig 12.b

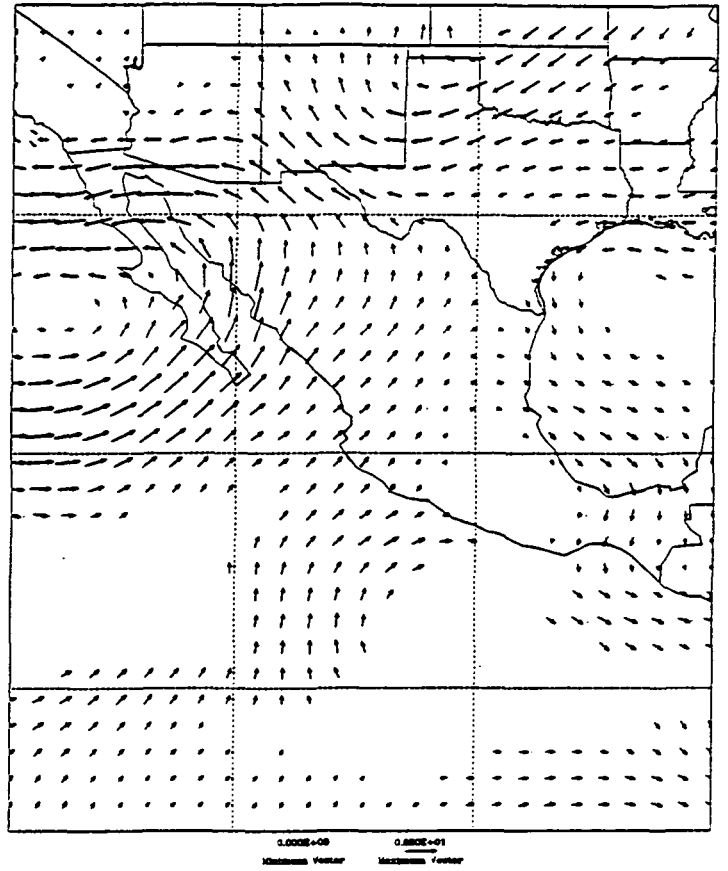


Fig 12.c

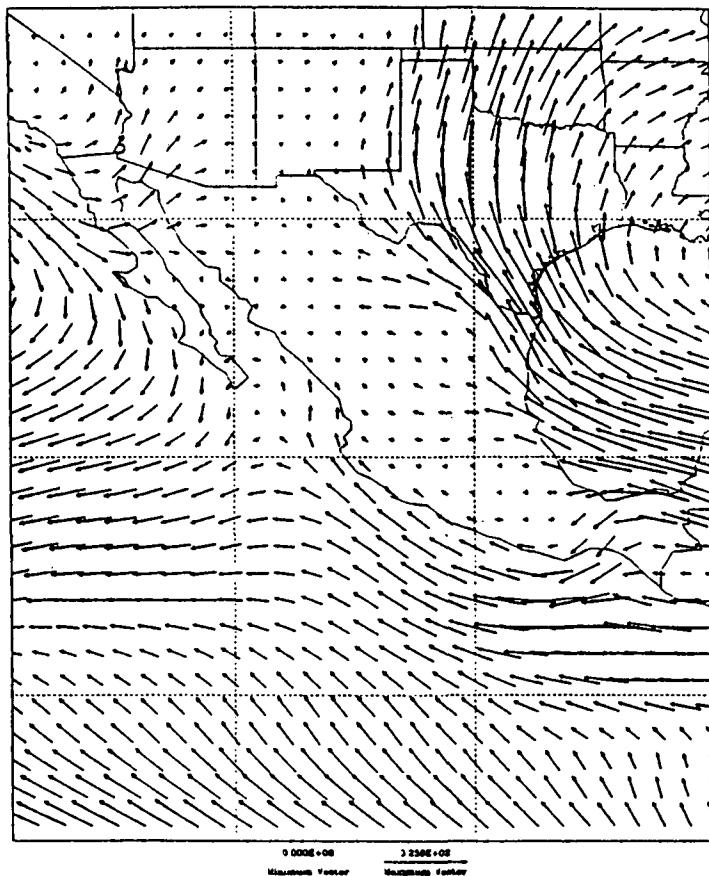


Fig 13.a

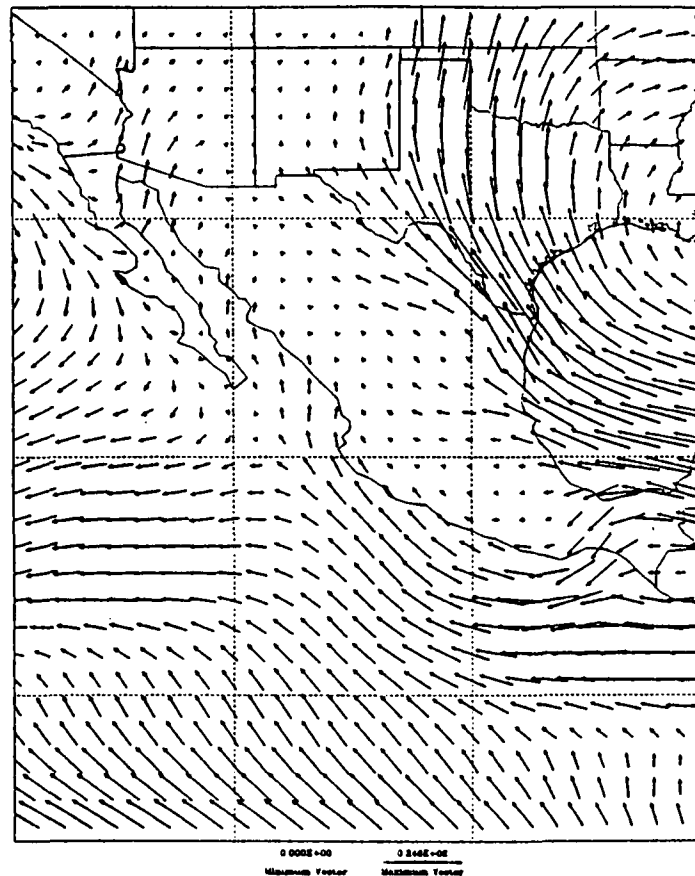


Fig 13.b

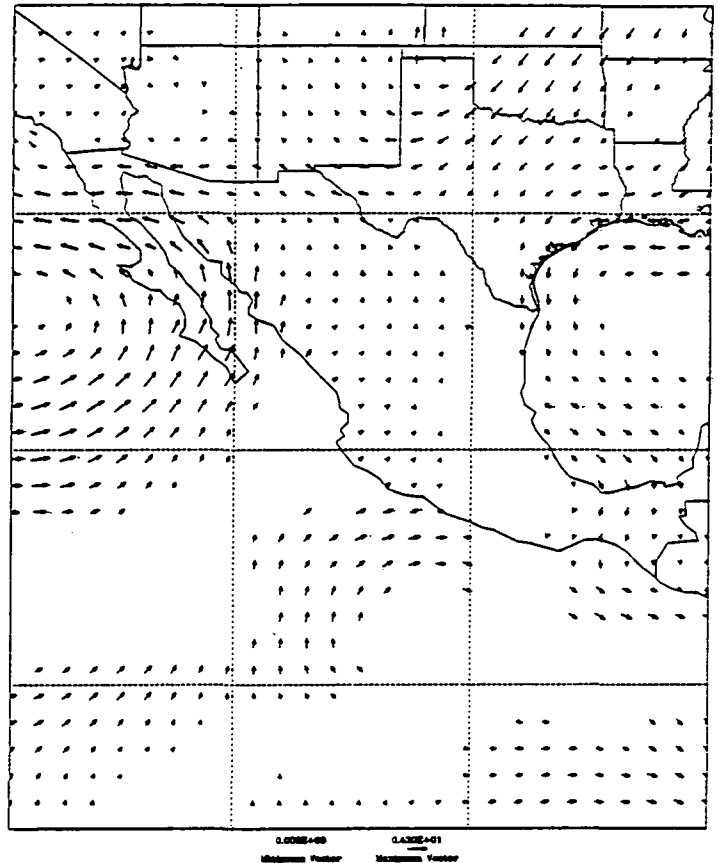
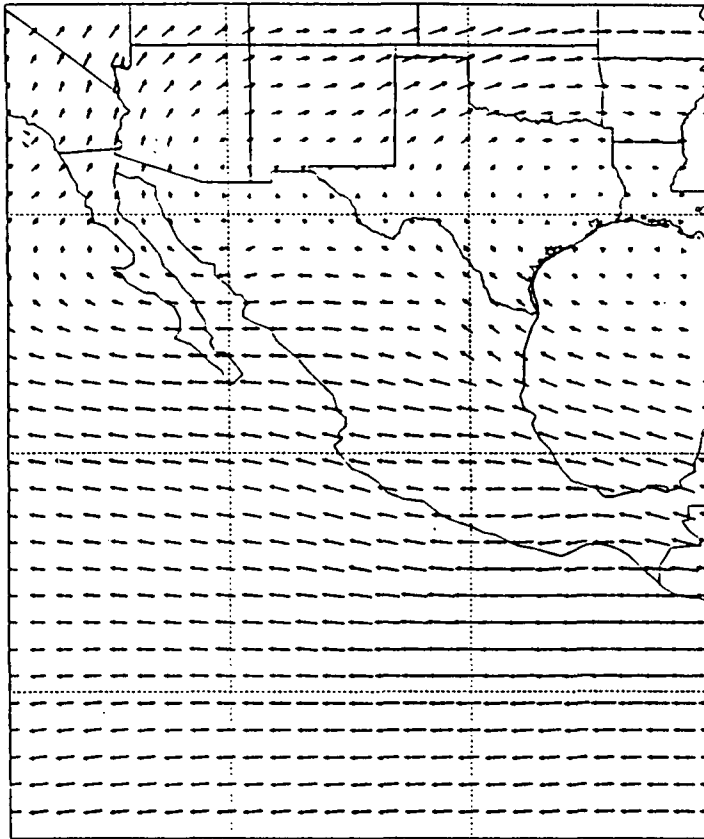
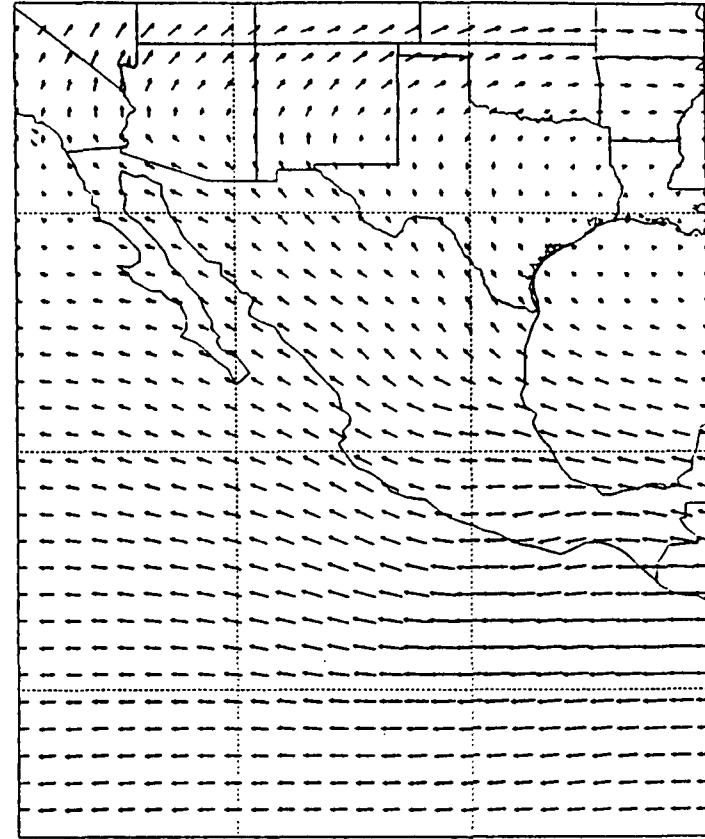


Fig 13.c



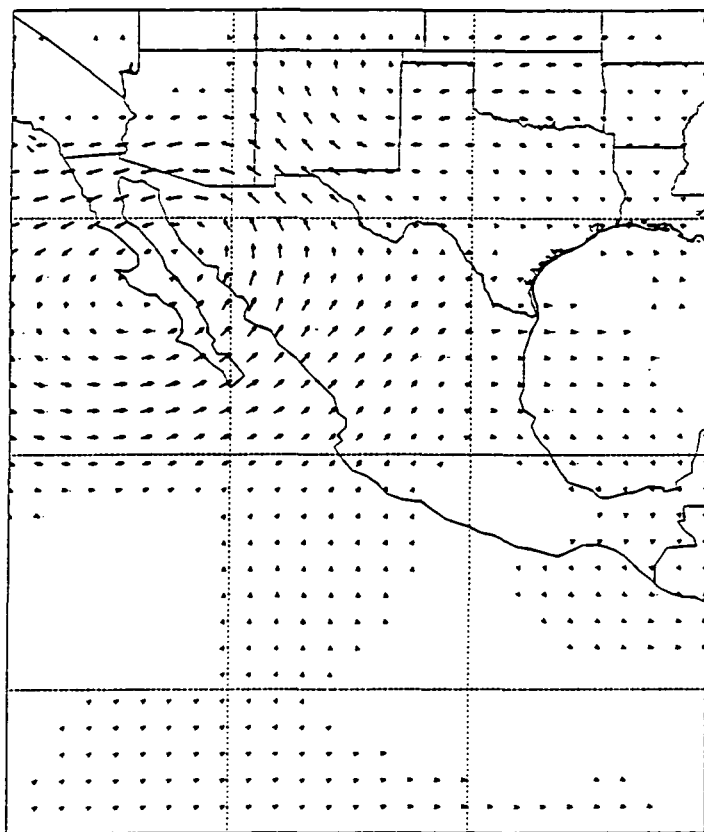
0.000E+00 0.000E+01
Minimum Vector Maximum Vector

Fig 14.a



0.000E+00 0.000E+01
Minimum Vector Maximum Vector

Fig 14.b



0.000E+00 0.000E+00
Minimum Vector Maximum Vector

Fig 14.c

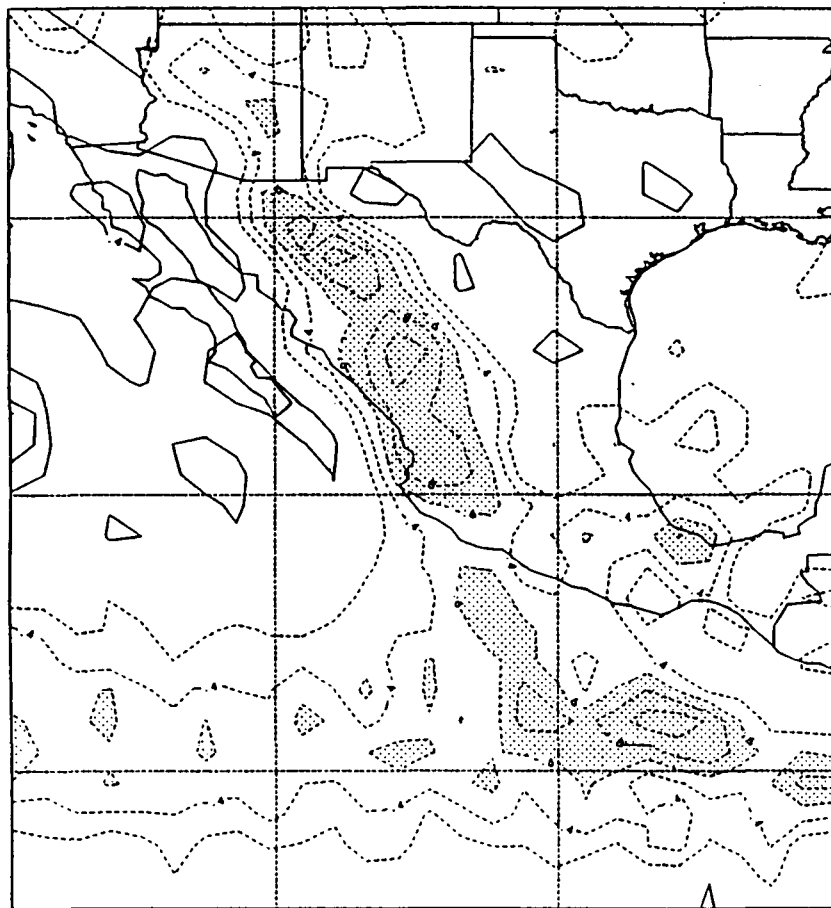


Fig 15.7

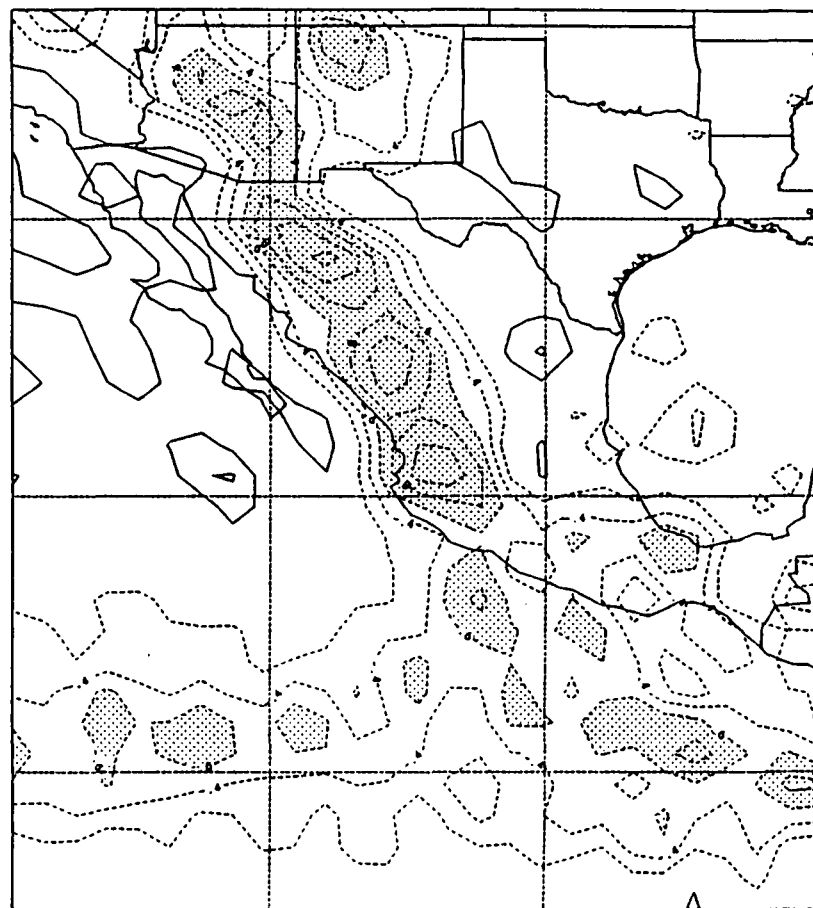


Fig 15.6

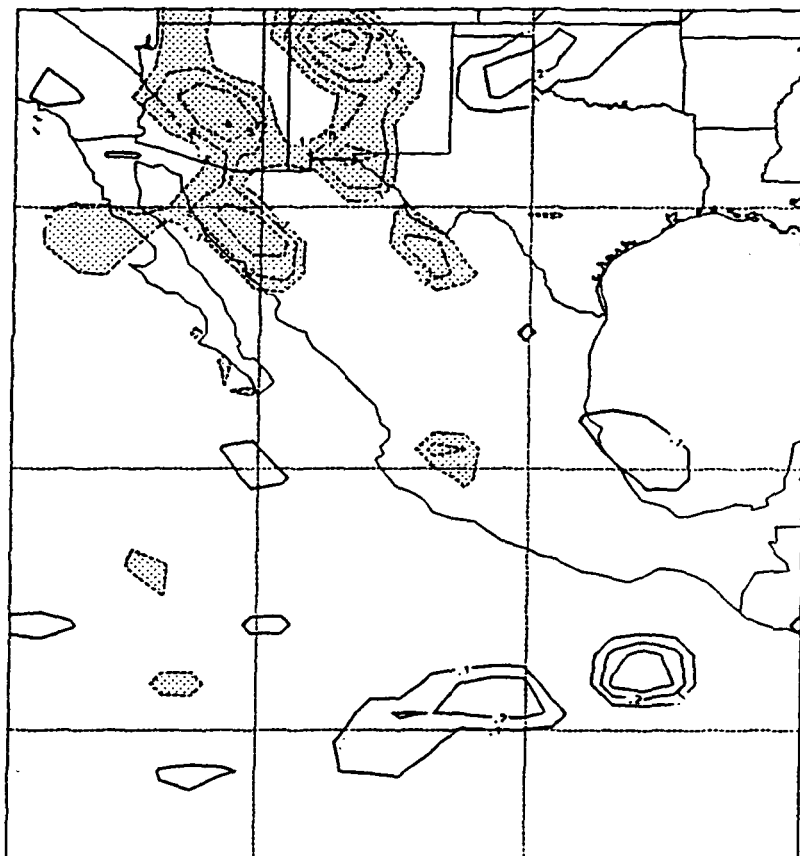


Fig 15.c

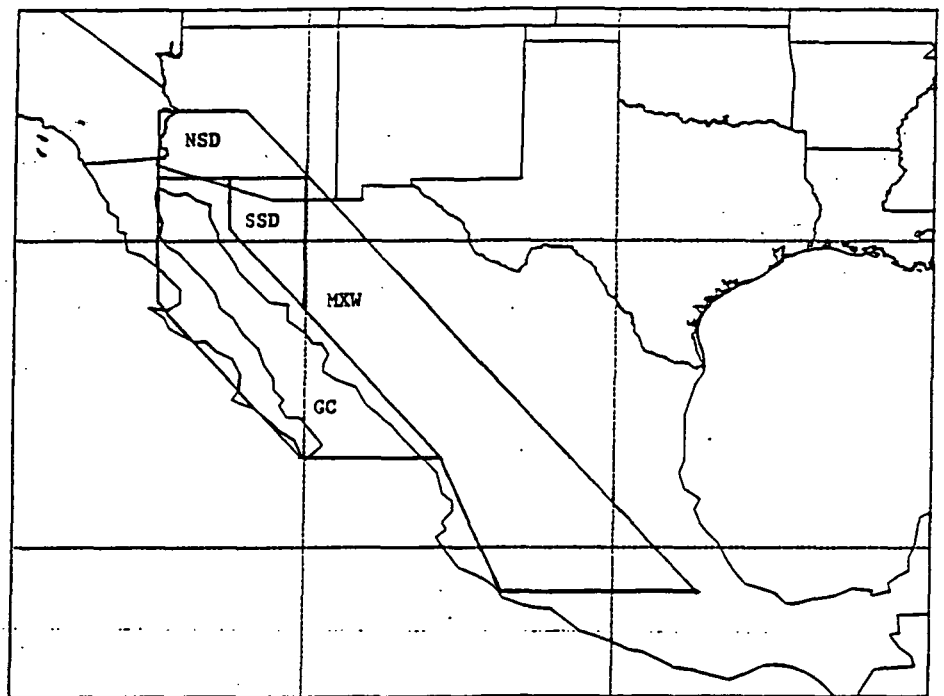


Fig 16

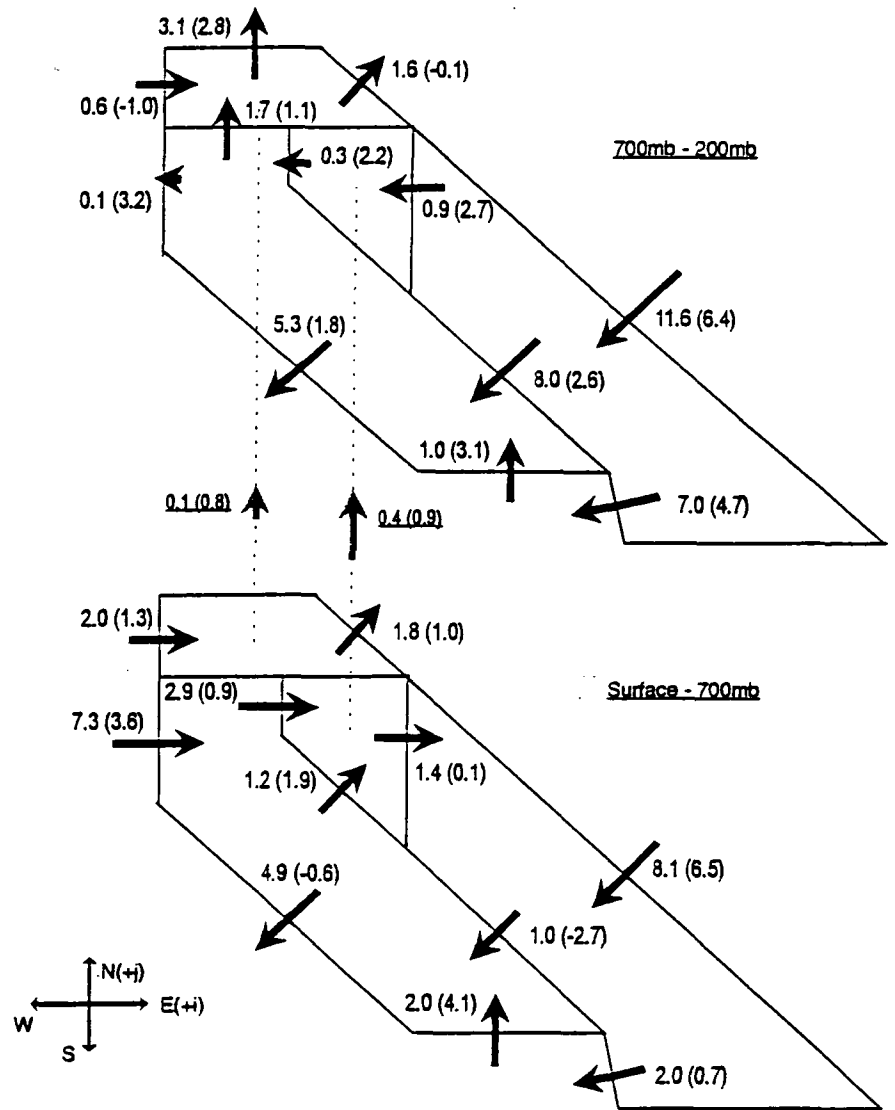


Fig 17

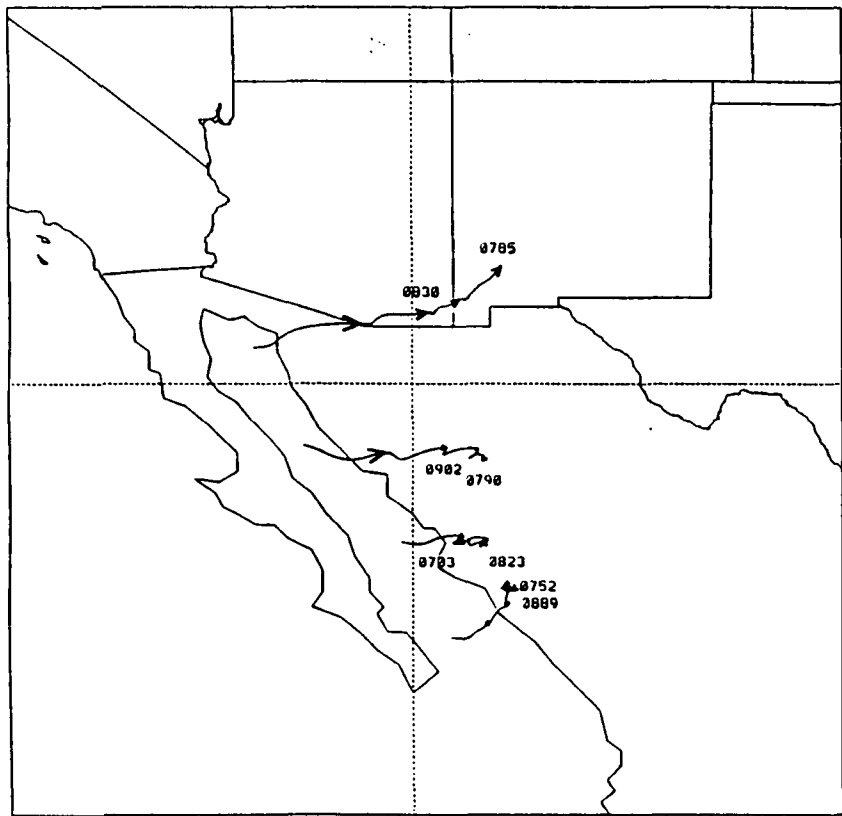


Fig 18.a

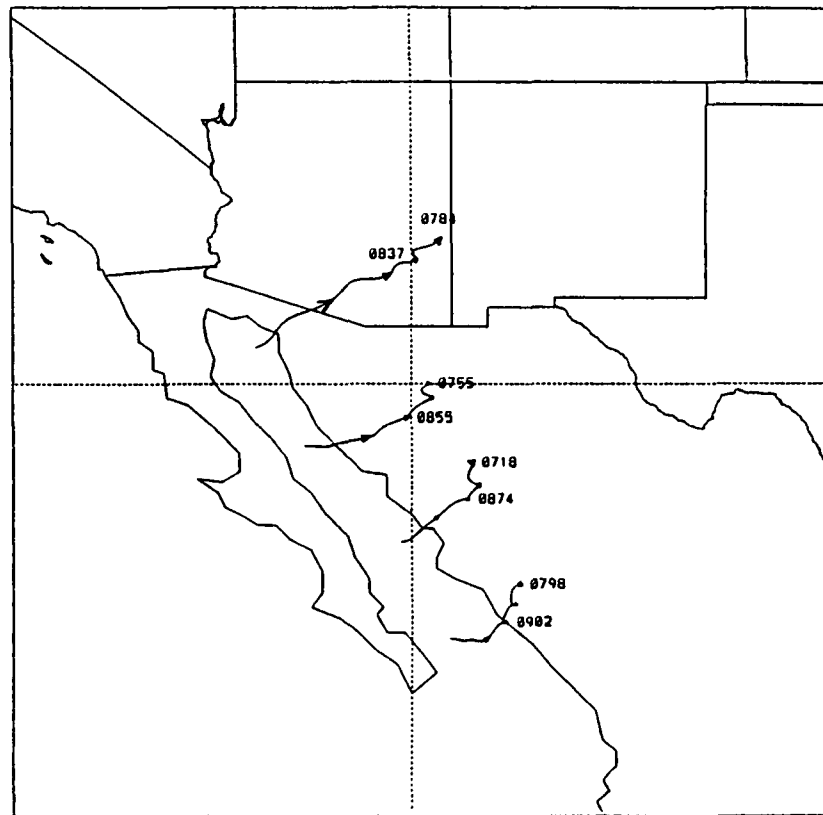


Fig 18.b

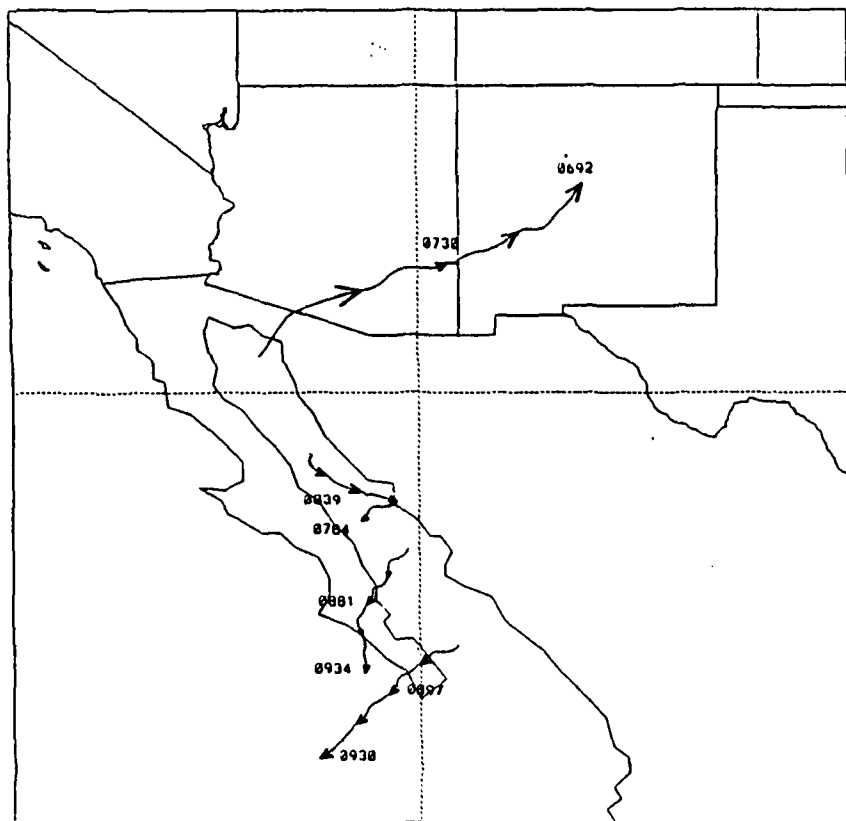


Fig 19.a

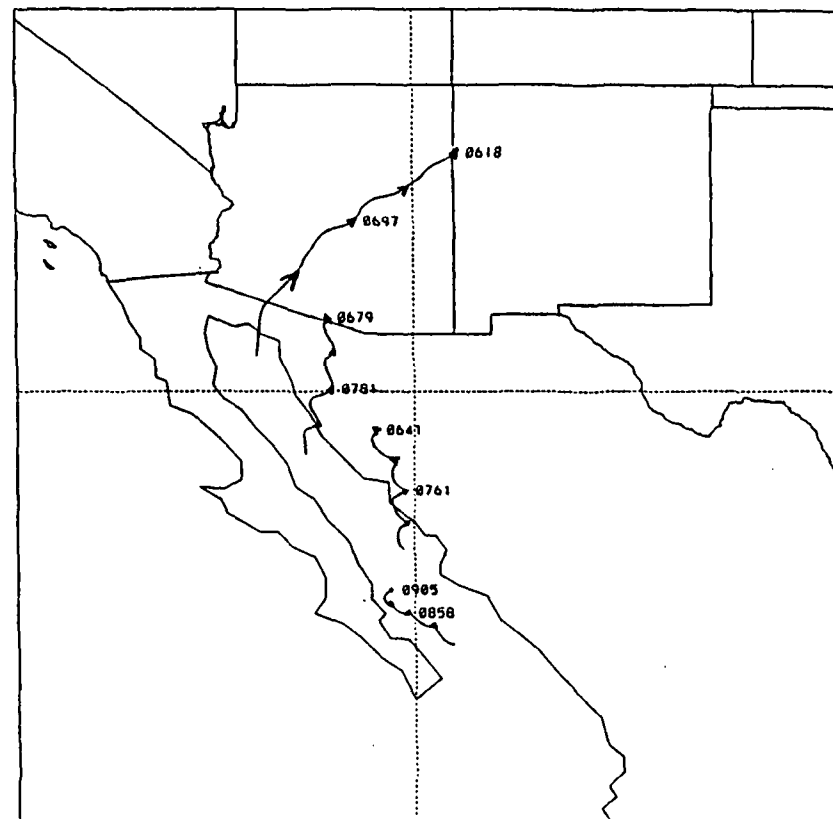


Fig 19.b

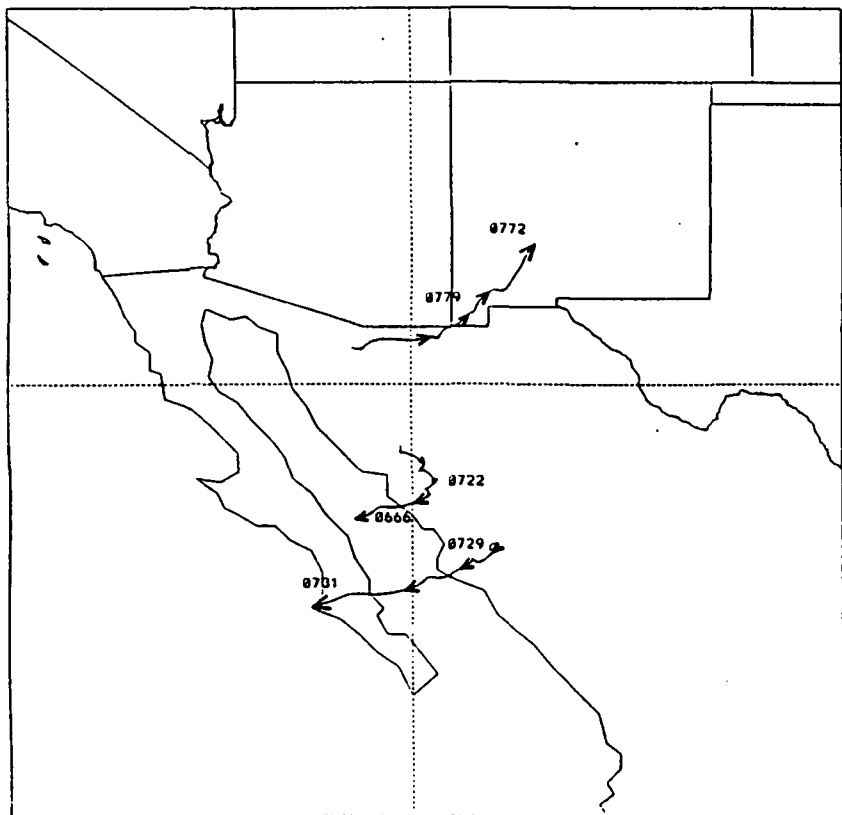


Fig 20.9

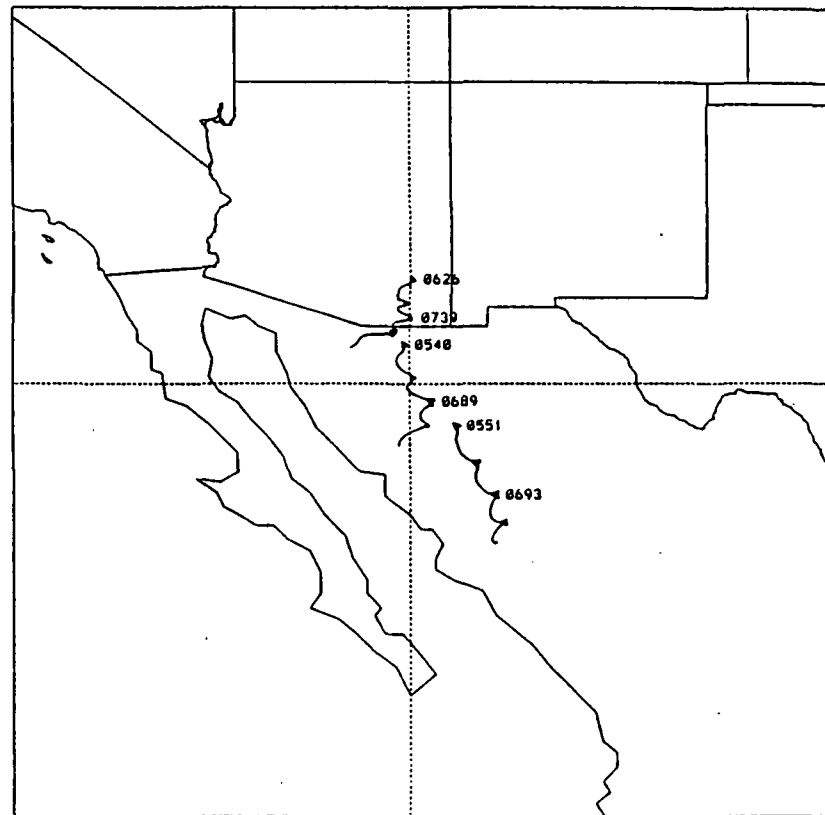


Fig 20.6

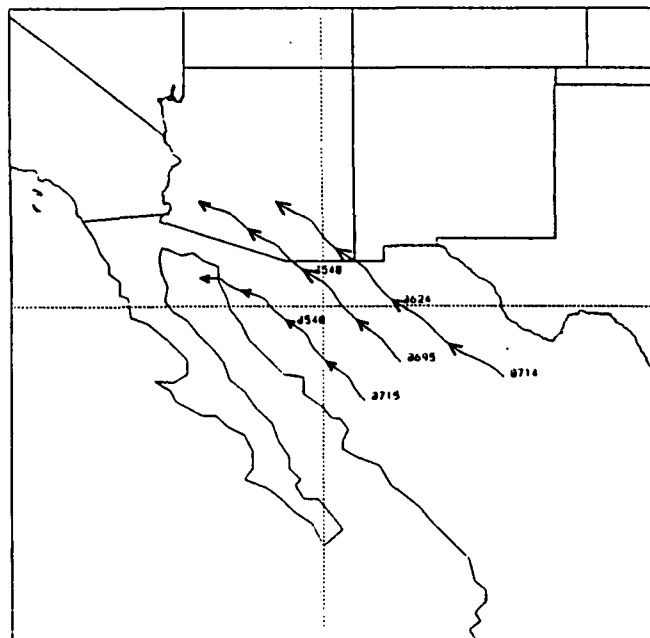


Fig 21.6

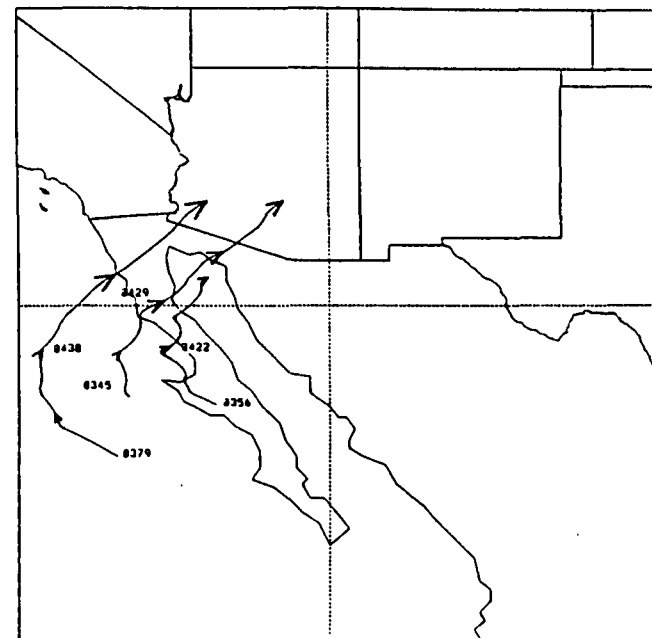


Fig 21.9